101-04613/

GEOLOGICAL SURVEY OF CANADA
BULLETIN 548

A SYNTHESIS OF GEOLOGICAL HAZARDS IN CANADA

G.R. Brooks (editor)

2001

©Her Majesty the Queen in Right of Canada, 2001

Catalogue No. M42-548E
ISBN 0-660-18316-1

Available in Canada from
Geological Survey of Canada offices:

601 Booth Street
Ottawa, Ontario K1A 0E8

3303-33rd Street N.W.
Calgary, Alberta T2L 2A7

101-605 Robson Street
Vancouver, B.C. V6B 5J3

A deposit copy of this publication is also available for reference
in selected public libraries across Canada

Price subject to change without notice

Cover illustration

Flooding in the Red River Valley, Manitoba, during 1997 (photograph by Greg Brooks (GSC 2000-073)). Damaged buildings, Lord's Cove, Newfoundland, 1929 tsunami (photo by Harris M. Mosdell, from the collection of W.M. Chisholm; provided by Alan Ruffman). Shoreline erosion at Tuktoyaktuk, N.W.T. despite efforts of stabilization (photograph courtesy of Steve Soloman (GSC 2000-074)). The Hope Slide, British Columbia , which occurred on January 9, 1965 (British Columbia Government aerial photograph 0445).

Editor's address
G.R. Brooks
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
K1A 0E8

Original manuscript submitted: 1998-12
Final version approved for publication: 2000-08

PREFACE

On 22 December 1989, the United Nations General Assembly adopted a resolution proclaiming the 1990s to be the International Decade for Natural Disaster Reduction (IDNDR). Canada was a co-sponsor of this resolution.

The idea for the Decade sprang from the observation that losses due to natural disasters were increasing rapidly. Worldwide, the number of catastrophes had quadrupled over a 20-year period. This led to the proposal of the Decade by Dr. Frank Press, former President of the United States National Academy of Sciences, who took the suggestion to the United Nations.

In support of this proposal, a Joint Committee of the Royal Society of Canada and the Canadian Academy of Engineering was formed in 1988; it endorsed the United States Academy of Sciences initiative and produced a report entitled *Toward a Canadian Program for the International Decade for Natural Disaster Reduction*, which recommended the formation of the Canadian National Committee for the International Decade for Natural Disaster Reduction with broad representation of experts with experience in natural disasters from both the public and private sectors. This committee was formed in 1993. Its first task was to prepare *The Canadian National Report* on natural disasters in Canada for the International Decade for Natural Disaster Reduction midterm review at the 1994 World Conference on Natural Disaster Reduction held in Yokohama, Japan. In this the Geological Survey of Canada (GSC) took a leading role.

With the threat of catastrophic losses, understanding geological hazards has become increasingly important. The GSC, in its over 150 years of existence, has made immense contributions to the economic and scientific knowledge of the country. This GSC volume on natural geological hazards is one more example of such a contribution.

One of the targets for the Decade set by the United Nations General Assembly is that, by the year 2000, all countries, as part of their plan to achieve sustainable development, should have in place comprehensive national assessments of risk from natural hazards, with these assessments taken into account in development plans.

The contributors to this outstanding GSC Hazard Synthesis bulletin have ensured that in these respects Canada meets the targets for the International Decade for Natural Disaster Reduction. This is a basic step in mitigating natural disasters in Canada.

Susan Till
Acting Assistant Deputy Minister
Natural Resources Canada

PRÉFACE

Le 22 décembre 1989, l'Assemblée générale des Nations Unies proclamait les années 1990 la Décennie internationale de prévention des catastrophes naturelles (DIPCN). Le Canada était un cosignataire de cette résolution.

Cette idée était venue après que l'on eut constaté une augmentation rapide des pertes attribuables aux désastres naturels. À l'échelle de la planète, le nombre de catastrophes avait quadruplé sur une période de vingt ans. La proposition fut donc présentée aux Nations Unies par M. Frank Press, ancien président de l'Académie nationale des sciences des États-Unis.

La Société royale du Canada et l'Académie canadienne du génie ont formé en 1988 un comité mixte pour appuyer cette proposition. Ce comité a souscrit à l'initiative de l'Académie nationale des sciences des États-Unis et produit un rapport intitulé *Vers un programme canadien pour la Décennie internationale de la prévention des catastrophes naturelles*. Le rapport recommandait la formation d'un Comité national canadien pour la Décennie composé d'un grand nombre d'experts du secteur privé et du secteur public représentant diverses disciplines et possédant de l'expérience dans le domaine des désastres naturels. Ce comité a été formé en 1993. Sa première tâche a été de préparer *Le Rapport national canadien* sur les

catastrophes naturelles au Canada, qui devait faire l'objet d'un examen de mi-parcours à la Conférence mondiale sur la prévention des catastrophes naturelles de 1994, à Yokohama, au Japon. La Commission géologique du Canada (CGC) a joué un rôle central dans ce projet.

Pour contrer la menace de pertes catastrophiques, il est de plus en plus important de comprendre les processus à l'origine des désastres naturels. En 150 années d'existence, la CGC a fait d'immenses contributions au savoir économique et scientifique de ce pays. Le volume de la CGC sur les risques naturels n'est qu'un exemple parmi beaucoup d'autres.

Un des objectifs de la Décennie proclamée par l'Assemblée générale des Nations Unies est qu'en l'an 2000 tous les pays devraient avoir procédé à des évaluations nationales complètes des risques de catastrophes naturelles et en tenir compte dans leurs plans d'aménagement et de développement durable.

Les participants à cette extraordinaire synthèse des risques de la CGC ont fait en sorte que le Canada respecte les objectifs de la Décennie internationale de la prévention des catastrophes naturelles. Il s'agit d'une première étape pour atténuer les impacts des catastrophes naturelles au Canada.

Susan Till
Sous-ministre adjointe
par intérim
Ressources naturelles Canada

CONTENTS

Preface/Préface iii

Summary/Sommaire
G.R. Brooks 1

Catastrophic hazards

Seismicity and seismic hazards
J. Adams and P. Basham 7

Tsunamis
J.J. Clague 27

Landslides
S.G. Evans 43

Snow avalanches
B. Jamieson 81

Floods
G.R. Brooks, S.G. Evans, and J.J. Clague 101

Volcanoes and volcanic hazards
C.J. Hickson and B.R. Edwards 145

Geomagnetic hazards
D.H. Boteler 183

Impact cratering on Earth
R.A.F. Grieve 207

Hazardous conditions

Sea-level rise in Canada
J. Shaw, R.B. Taylor, D.L. Forbes, and S. Solomon 225

Environmental geochemistry and geochemical hazards
R.D. Knight and R.A. Klassen (compilers) 227

Eolian activity
S.A. Wolfe 231

Permafrost in Canada, a challenge to northern development
S.L. Smith, M.M. Burgess, and J.A. Heginbottom 241

Natural gas hydrates
S.L. Smith 265

Author index 281

SUMMARY

G.R. Brooks

The Canadian landscape is prone to naturally occurring geological hazards that affect our society. There are many examples of settlement or economic development occurring with an ignorance or underestimation of local hazards, only to lead to disaster in the next hazard event or significant impacts from a slow-onset hazard. Such problems arise in part because Canada is relatively young compared to old-world countries and, therefore, our collective experience of geological hazards is limited. However, geological hazards can be recognized and understood through scientific research, and their impacts minimized through avoidance or mitigation. It is essential that hazards research span provincial and municipal borders to obtain as complete an understanding of a given phenomenon as possible, because the occurrence of geological hazards is controlled by local or regional factors, such as geology, geomorphology, and climate.

The need to study geological hazards is as great now as it has ever been. Globally, over the past several decades, there has been a marked increase in both the frequency and cost of major natural disasters from geological, hydrological, and meteorological causes. Canada has not been immune from this trend, much to the concern of the insurance industry and government. A number of reasons are behind this, including increased concentrations of population in disaster-prone areas, increased dependence of society on technology that is vulnerable to natural hazards, and increased economic prosperity creating more wealth that is vulnerable to damage. An additional factor is global climate change, which may affect the frequency and magnitude of many types of hazards.

As a national research agency, the Geological Survey of Canada (GSC) has investigated natural geological hazards since its inception in 1842, beginning with a description of a landslide in the Ottawa area by Sir William Logan (first director of the GSC). The study of geological hazards is currently an important part of GSC activities and covers a range of topics throughout Canada as part of the GSC's Hazards and Environmental Geology Program. This program is driven by a concern for public safety, the protection of the environment, and the need for sustainable development. As Canada's premier geological institute, the GSC draws upon its pool of expertise and facilities for hazard research. The research is carried out by all GSC scientific divisions and involves scientists from a broad range of disciplines.

This bulletin presents a synthesis of the geological hazard topics addressed by the GSC, but also includes snow avalanches as noted below. It is intended to provide background information to readers and to direct them to more detailed sources of information. Each contribution consists of a succinct summary of a given topic that is written for a general rather than a

SOMMAIRE

G.R. Brooks

Le paysage canadien est enclin aux risques naturels qui affectent notre société. Il existe de nombreux exemples de peuplements ou de développement économique qui se font en ignorant ou en mésestimant les risques locaux, ce qui entraîne des catastrophes dans le cas de risques naturels ou des incidences considérables lorsqu'il s'agit de risques à déclenchement lent. Ces problèmes surviennent en partie parce que le Canada est un pays relativement jeune en comparaison des pays du vieux continent et, par conséquent, notre expérience collective des risques naturels est limitée. Cependant, la recherche scientifique permet de reconnaître et de comprendre ces risques naturels et de minimiser leurs impacts en les évitant ou en les atténuant. Afin de comprendre un phénomène donné de manière la plus exhaustive possible, il est essentiel que les travaux de recherche sur les risques naturels chevauchent les frontières provinciales et municipales, car la venue de risques naturels est contrôlée par des facteurs locaux ou régionaux tels que la géologie, la géomorphologie et le climat.

L'étude des risques naturels est aussi importante de nos jours qu'elle ne l'était dans le passé. À l'échelle planétaire, au cours des dernières décennies, la fréquence et le coût des grandes catastrophes naturelles de nature géologique, hydrogéologique et météorologique ont augmenté de manière sensible. Le Canada n'a pas échappé à cette tendance, ce qui constitue un sujet de préoccupations pour les sociétés d'assurance et le gouvernement. Il existe plusieurs raisons à cet état de fait, notamment les concentrations plus grandes de population dans des régions à risque, la dépendance accrue de la société sur une technologie qui est sensible aux risques naturels et une plus grande prospérité économique créatrice d'une plus grande richesse, laquelle est vulnérable aux avaries. Un autre facteur, le changement du climat de la planète, est susceptible d'influer sur la fréquence et l'ampleur de nombreux types de risques.

À titre d'organisme de recherche national, la Commission géologique du Canada (CGC) étudie les risques naturels depuis sa création en 1842. La première étude, effectuée par Sir William Logan (premier directeur de la CGC), décrivait un glissement de terrain dans la région d'Ottawa. De nos jours, l'étude des risques naturels constitue une partie importante des activités de la CGC et porte sur un éventail de sujets faisant l'objet de recherche à l'échelle du Canada dans le cadre du Programme sur les risques naturels et la géologie de l'environnement. Ce programme est animé par le souci d'assurer la sécurité publique, la protection de l'environnement et le développement durable. À titre de premier institut de géologie du Canada, la CGC met son vivier de spécialistes et ses nombreuses installations à la disposition de la recherche sur les risques naturels. Les travaux de recherche sont exécutés par toutes les divisions scientifiques de la CGC et mettent à contribution des scientifiques de disciplines très variées.

Le présent bulletin présente une synthèse des sujets relatifs aux risques naturels traités à la CGC, ainsi que des avalanches de neige comme il est précisé ci-dessous. Il a pour but de fournir aux lecteurs des renseignements de base et de les diriger vers des sources d'information plus élaborées. Chaque article comporte un sommaire du sujet traité et a été rédigé à l'intention du grand public plutôt que des spécialistes. Afin d'illustrer leurs propos, certain

specialized audience. Where appropriate and available, some contributions contain compilations of significant events and/or selected case studies. The bulletin was conceived to help raise awareness and increase the understanding of natural geological hazards; it also represents a contribution by the GSC to the International Decade for Natural Disaster Reduction sponsored by the United Nations.

Two basic types of hazards are distinguished in this bulletin, catastrophic hazards and hazardous conditions. Catastrophic or 'sudden-onset' hazards occur rapidly and with little or no warning. They represent obvious hazards and can result in human fatalities and damage to property and infrastructure. The catastrophic hazards summarized are earthquakes, tsunamis, volcanoes, meteorites, landslides, snow avalanches, geomagnetic hazards, and floods. The definition of 'sudden-onset' can be moot in some specific cases of a given phenomenon. For example, some types of landslides occur as creep, geomagnetic activity can result in the slow corrosion of a pipeline, eventually leading to a blowout, and some types of floods are predictable in the short term and can rise and fall slowly over a period of many weeks. Nevertheless, in general, all these hazards can occur rapidly within seconds, minutes, or hours, or over several days.

Hazardous conditions, or 'slow-onset' hazards, are more subtle phenomena that occur relatively slowly and do not immediately threaten lives; however, they can cause property damage or public health problems over the longer term (months, years). Hazardous conditions occur regionally and must be taken into account when conducting vulnerable activities or establishing infrastructure to avoid problems later. The hazardous conditions included here are sea-level rise, permafrost, geochemical activity in the environment, eolian processes, and gas hydrates. Some of these phenomena can lead to catastrophic impacts. For example, a disturbance to permafrost can lead to landsliding, and severe wave actions during storms can result in major shoreline erosion in areas where sea level is rising gradually. Overall, however, the basic notion that hazardous conditions exist in the background or operate at a low intensity holds true.

The geological hazard topics are presented as individual contributions, written by GSC scientists. Two of the contributions ('Sea-level rise' and 'Environmental geochemistry') are contained as pocket posters. All contributions are organized under the general headings of 'Catastrophic hazards' and 'Hazardous conditions'.

Except for snow avalanches, all the hazard topics contained in this bulletin are the subject of GSC research. The classification of snow avalanches is problematic because although the material (snow) is the product of meteorological processes, it undergoes

auteurs présentent des compilations d'événements importants ou des études de cas. Ce bulletin vise à faire prendre connaissance des risques naturels et se veut être une contribution de la CGC à la Décennie internationale de la prévention des catastrophes naturelles parrainée par les Nations Unies.

Deux types fondamentaux de risques naturels sont présentés, soit les risques catastrophiques et les conditions dangereuses. Les risques catastrophiques ou «à déclenchement soudain» se produisent brusquement avec peu ou pas de signes avant-coureurs. Ils représentent des risques évidents et peuvent provoquer des pertes de vies humaines et des dommages aux biens et aux infrastructures. Dans le cadre de la présente synthèse, les risques catastrophiques étudiés sont les séismes, les tsunamis, les volcans, les météorites, les glissements de terrain, les avalanches de neige, les orages magnétiques et les inondations. La définition de «déclenchement soudain» peut avoir un caractère spéculatif dans le cas d'un phénomène particulier. Ainsi, certains types de glissement de terrain se produisent sous la forme de reptation, une activité géomagnétique peut entraîner la corrosion lente d'un pipeline pouvant conduire à sa rupture, et certains types d'inondation prévisibles à court terme peuvent survenir et disparaître lentement sur une période de nombreuses semaines. Cependant, en règle générale, tous ces risques peuvent se déclencher en quelques secondes, minutes ou heures ou s'échelonner sur plusieurs journées.

Les conditions dangereuses ou «à déclenchement lent» sont des phénomènes plus discrets qui se produisent plutôt lentement et qui ne mettent pas immédiatement des vies en danger. Cependant, ils peuvent endommager les biens ou créer des problèmes de santé publique à long terme (mois et années). Ces conditions se produisent à l'échelle régionale et pour éviter des problèmes dans l'avenir, elles doivent être prises en compte lors de l'exécution d'activités potentiellement à risques ou de l'établissement d'infrastructures. Les conditions dangereuses traitées dans ce bulletin comprennent la montée du niveau de la mer, le pergélisol, l'activité géochimique dans l'environnement, les processus éoliens et les hydrates de gaz. Certains de ces phénomènes peuvent entraîner des incidences catastrophiques. Ainsi, la perturbation du pergélisol peut déclencher des glissements de terrain et l'action violente des vagues au cours de tempêtes peut entraîner l'érosion importante des côtes dans des régions où le niveau de la mer monte progressivement. En règle générale, la notion fondamentale selon laquelle les conditions dangereuses sont latentes ou agissent à base intensité demeure une vérité.

Les sujets portant sur les risques naturels sont présentés sous forme d'article individuel rédigé par des scientifiques de la CGC. Deux articles (*Sea-level rise* et *Environmental geochemistry*) sont présentés sous forme d'affiches en pochette. L'ensemble des publications sont réunies sous les deux rubriques *Catastrophic hazards* («Risques catastrophiques») et *Hazardous conditions* («Conditions dangereuses»).

À l'exception des avalanches de neige, tous les sujets traités dans ce bulletin font partie de travaux de recherche effectués à la CGC. Le classement des avalanches de neige présente des difficultés, car bien que la matière (la neige) soit le produit de processus météorologiques, elle est l'objet de perturbations

geotechnical failures analogous to some types of landslides. As a result, both the meteorological and geological communities commonly ignore the topic. A paper on snow avalanches is included for the sake of completeness, because snow avalanches are a very significant hazard in Canada and it closely complements the contribution on landslides.

The monitoring and study of earthquakes is a major component of the GSC hazard program. Earthquakes occur along Canada's western margin as well as in the Arctic and southeastern regions. Hazards associated with earthquakes include shaking, ground rupture and land-level change, tsunamis, sediment liquefaction, and landslides. Historical earthquakes have damaged buildings and other structures, primarily though foundation failures from shaking. To aid in the design of earthquake-resistant buildings, the GSC is producing a fourth-generation seismic hazard map to assign updated seismic hazard levels across the country.

Tsunamis are waves produced by earthquakes and, less commonly, by landslides, volcanic eruptions, and meteorite impacts. The coast of British Columbia is struck by large earthquake-generated tsunamis on average once every few hundred years. Tsunamis caused by landslides are much more localized, but can occur in lakes or fiords. Tsunamis are uncommon on the Atlantic and Arctic coasts. Although tsunamis cannot be prevented, their impacts can be reduced through emergency preparedness, public education and a variety of nonstructural and structural measures.

Landslides are mass movements of geological materials and range in size from an individual boulder to large rock masses that form a mountainside. They occur in different geological materials, display a variety of movement modes, and move extremely rapidly to very slowly. They are important geomorphic processes in many regions of Canada and have caused considerable property damage. Between 1840 and 1998, there have been 43 landslide disasters in Canada that have killed three or more people. There are many examples of landslide problems in Canada being successful mitigated. Since the balance of forces that control the stability of slopes are very sensitive to alterations in the immediate environment, landslide activity can be triggered by, for example, severe climatic events, anthropogenic disturbance, and seismic shaking.

Snow avalanches occur in many areas of Canada, including the Cordillera, the Maritime Provinces, and the Arctic, but mostly in remote or uninhabited regions. They represent a widespread, but in some areas an underappreciated, hazard. Over 570 people have died in snow avalanches in Canada since the mid-1800s. Currently, about 12 people die each year in avalanches. As well, traffic delays and damage to forests caused by avalanches cost millions of dollars annually in direct and indirect costs. The formation of snow avalanches is

géotechniques semblables à certains types de glissements de terrain. De ce fait, les météorologues et les géologues évitent de traiter ce sujet. À des fins d'exhaustivité, les avalanches de neige sont incluses dans ce bulletin, car elles constituent un risque majeur au Canada et sont un complément à l'article sur les glissements de terrain.

La surveillance et l'étude des séismes est une composante majeure du programme des risques mis sur pied par la CGC. Au Canada, les séismes se déclenchent le long de la marge occidentale du pays ainsi que dans les régions de l'Arctique et du sud-est. Les risques qui y sont associés sont notamment les secousses, les fissures et les dénivellations des terrains, les tsunamis, la liquéfaction des sédiments et les glissements de terrain. Les séismes historiques ont endommagé des constructions et d'autres structures, principalement en provoquant des accidents de fondations lors des secousses. Dans le but de contribuer à la conception de constructions parasismiques, la CGC dresse actuellement une carte des risques sismiques de quatrième génération afin d'attribuer à l'ensemble des régions du Canada des niveaux de risques sismiques mis à jour.

Les tsunamis sont des ondes engendrées par les séismes et, moins fréquemment, par les glissements de terrain, les éruptions volcaniques et les chutes de météorites. Le littoral de la Colombie-Britannique est touché à des intervalles moyens de quelques centaines d'années par d'importants tsunamis d'origine sismique. Les tsunamis causés par les glissements de terrain sont plus localisés, mais peuvent se produire dans les lacs ou les fjords. Les tsunamis se produisent rarement sur les côtes de l'Atlantique et de l'Arctique. Bien qu'il soit impossible de les prévenir, on peut réduire leurs répercussions en prévoyant des mesures d'urgence, en sensibilisant le public et en prenant des mesures structurelles et non structurelles.

Les glissements de terrain sont des déplacements de masses de matériaux géologiques dont la dimension varie d'un bloc simple à d'énormes masses rocheuses formant le flanc d'une montagne. Ils se produisent dans divers matériaux géologiques, selon des modes variés de déplacement, et se déplacent extrêmement vite ou très lentement. Dans certaines régions du Canada, ils constituent d'importants processus géomorphologiques et ont causé des dommages matériels considérables. Entre 1840 et 1998, le Canada a été le siège de 43 glissements de terrain destructeurs qui ont chacun causé la mort d'au moins trois personnes. Il existe de nombreux exemples de problèmes liés aux glissements de terrain qu'on a réussi à atténuer au Canada. Puisque l'équilibre des forces qui contrôlent la stabilité des pentes est très sensible aux modifications apportées à l'environnement immédiat, des conditions climatiques rigoureuses, des perturbations anthropiques et des secousses sismiques, par exemple, peuvent déclencher un glissement de terrain.

Les avalanches de neige se produisent dans plusieurs régions du Canada, notamment dans la Cordillère, les provinces maritimes et l'Arctique, mais principalement dans les régions éloignées ou inhabitées. Elles représentent un risque très répandu mais mésestimé dans certaines régions. Depuis le milieu du XIX^e siècle, plus de 570 personnes ont perdu la vie dans des avalanches de neige au Canada. De nos jours, environ 12 personnes meurent chaque année dans les avalanches. En outre, les entraves à la circulation et les dommages causés aux forêts par les avalanches entraînent chaque année des coûts directs et indirects de l'ordre de millions de dollars. Le déclenchement d'avalanches de neige est fortement liée

strongly influenced by slope angle and orientation to wind. Forecasting and control programs can reduce snow-avalanche hazards in ski areas and along transportation corridors. As well, avalanche hazards can be mitigated by using structural methods.

Flooding is a widespread and common hazard in Canada that has caused 168 known Canadian flood 'disasters' between 1900 and June 1997. Although representing a hazard to communities and structures that encroach upon flood-prone lands, floods are an integral part of the flow regime of rivers. Flood impacts can be reduced by the careful management of flood-prone lands and by the construction of flood protection infrastructure. Floods cannot be predicted over long periods, but for some watersheds, outlooks of peak spring discharge can be forecast weeks in advance, and flood advisories or warnings can be issued often hours in advance of expected severe weather conditions. Climate change may alter the frequency of extreme flood events, thus increasing the risk of occupying flood-prone lands along some rivers.

Volcanoes in Canada are found in British Columbia and the Yukon Territory, primarily within five volcanic fields. Although none of the volcanoes are active, some have erupted in the very recent past. Volcanoes in Canada have different morphologies and scales and range from cinder cones to stratovolcanoes and shield volcanoes. Interestingly, a number of the volcanoes have erupted beneath or immediately adjacent to ice during past glaciations yielding distinctive morphologies and/or rock types. Magma composition controls how a volcano erupts. Low-viscosity magmas erupt passively whereas high-viscosity magmas erupt explosively. Although all volcanic eruptions represent an obvious hazard to areas immediately adjacent to the volcano, explosive eruptions are far more hazardous, as they can produce pyroclastic flows and thick clouds of fine ash. Because of the great potential travel distance of volcanic ash clouds, Canada can be directly effected by large explosive eruptions of volcanoes located in the United States.

Over the past 150 years, a number of electrical conducting networks have been constructed over large parts of the globe, including telecommunication cables, high-voltage transmission lines, and pipelines. All these technological systems are enveloped by the Earth's magnetic field and will be affected by electric fields induced by changes of the Earth's magnetic field. When disturbances on the sun cause magnetic storms, the variations in the Earth's magnetic field become extreme and the induced electric fields can have a significant effect on power systems, pipelines, and telephone cables. Severe geomagnetic disturbances occur in cycles of approximately 11 years that coincide with increases in the number of sunspots on the sun's surface.

Of the approximately 160 known terrestrial impact structures distributed worldwide, 29 are located in Canada, mostly on the Canadian Shield. The basic

à l'angle de la pente et l'orientation de celle-ci par rapport aux vents. Les programmes de prévision et de contrôle peuvent réduire les risques d'avalanche dans les régions où l'on pratique le ski et le long des voies de communication. Les risques d'avalanche peuvent également être atténués en appliquant des méthodes structurelles.

Les inondations constituent un risque répandu et commun au Canada et ont été la cause de 168 «catastrophes» recensées entre 1900 et juin 1997. Bien qu'elles représentent un risque pour les populations et les structures qui empiètent sur les terrains inondables, les inondations font partie intégrante du régime d'écoulement des rivières. Une gestion judicieuse des terrains inondables et la construction d'infrastructures de défense contre les crues peuvent réduire les conséquences néfastes des inondations. S'il est impossible de prévoir les inondations à long terme, on peut prévoir plusieurs semaines à l'avance l'arrivée des crues printanières dans certains bassins hydrologiques et émettre, souvent plusieurs heures à l'avance, des avis ou des avertissements sur la venue d'inondations liées à la prévision de conditions climatiques particulièrement mauvaises. Le changement climatique est susceptible de modifier la fréquence des grandes inondations, augmentant ainsi le risque de construire sur les terrains inondables le long de certaines rivières.

Au Canada, les volcans sont situés en Colombie-Britannique et dans le Territoire du Yukon, principalement dans cinq champs volcaniques. Bien qu'aucun ne soit actif, certains volcans ont fait éruption dans un passé très récent. La forme et la dimension des appareils volcaniques du Canada varient, allant de cônes de scories à des stratovolcans et à des volcans-boucliers. Fait intéressant à noter, un certain nombre de volcans ont fait éruption sous des glaciers ou tout près au cours de glaciations passées, produisant des morphologies ou des types de roches caractéristiques. Le mode d'éruption du volcan est lié à la composition du magma. L'éruption d'un magma très fluide sera passive alors que celle d'un magma très visqueux sera explosive. Bien que toutes les éruptions volcaniques représentent un risque évident pour les régions immédiatement attenantes au volcan, les éruptions explosives sont, et de loin, les plus dangereuses puisqu'elles éjectent des coulées de débris pyroclastiques et d'épais nuages de cendres fines. En raison de la grande distance que peuvent parcourir les nuages de cendres volcaniques, le Canada peut être directement touché par de fortes éruptions explosives de volcans situés aux États-Unis.

Au cours des 150 dernières années, un certain nombre de réseaux conducteurs d'électricité ont été construits sur de vastes régions du globe, notamment des câbles de télécommunication, des lignes de transmission de haute tension et des pipelines. Tous ces systèmes techniques sont soumis à l'influence du champ magnétique terrestre et seront affectés par les champs électriques qui sont induits par les variations du champ magnétique terrestre. Les perturbations qui se produisent sur le soleil en raison des orages magnétiques entraînent des variations extrêmes dans le champ magnétique terrestre, et les champs électriques induits peuvent affecter considérablement les réseaux électriques, les pipelines et les câbles téléphoniques. Les perturbations magnétiques de forte intensité sont cycliques et se produisent environ tous les 11 ans. Elles coïncident avec l'accroissement du nombre de taches solaires sur la surface du soleil.

Des quelque 160 astroblèmes terrestres recensés de par le monde, 29 sont situés au Canada, la plupart dans le Bouclier canadien. Les types de base sont les cratères «simples», faisant jusqu'à 4 km

types of impact structures are 'simple' craters, up to 4 km in diameter, with uplifted and overturned rimrocks surrounding a bowl-shaped depression, and 'complex' structures, generally greater than 3 km in diameter, with a central uplift in the form of a peak and/or ring, an annular trough, and a slumped rim. The confirmation of the origin of an impact structure, however, is not based on morphology. Smaller impact craters are identified from the presence of fragments of the impactor. Larger impact craters are identified from a distinctive deformation within rocks known as 'shock metamorphism'. The formation of impact craters as small as 20 km can throw dust into the atmosphere that reduces sunlight and disrupts temperature affecting the climate and the biosphere to varying degrees at a global scale. Impacts into oceans can generate tsunamis that can be considerably larger than those generated by earthquakes. A major impact event at the Cretaceous–Tertiary (K–T) boundary is increasingly accepted as the cause of mass animal extinctions that occurred at that time.

Sea levels along Canada's coasts have fluctuated for thousands of years, primarily due to glacio-isostasy and eustasy. The combination of glacio-isostasy and eustasy is producing patterns of sea-level change that vary both spatially and temporally along the coasts. In areas where sea level is falling, for example in Hudson Bay, the coastline has retreated slowly seaward, so that old beaches can be found many metres above the modern coast and many kilometres inland. Many areas where sea level is rising, such as the east coast of Nova Scotia, are vulnerable to coastal erosion and storm surge, particularly during extreme storms. Coastal erosion can cause the loss of property and infrastructure, and flooding from storm surges can damage agricultural land, property, and infrastructure. The present pattern of sea-level change along Canadian coasts may be significantly altered over the next century due to the thermal expansion of ocean waters and melting of the Greenland and Antarctic ice caps, all arising from the effects of climate change.

Environmental geochemistry is the study of the abundance, distribution, and mobility of elements in bedrock and in the gravel, sand, silt, and clay sediments that form the Earth's surficial deposits. A wide range of geochemical properties in our landscape reflect variations in the composition of surficial materials and bedrock as well as geomorphic history. Environmental geochemistry is presented here from a geological perspective and thematic aspects of the subject that relate to hazards are summarized. Emphasis is placed on a small group of elements that, in abnormal concentrations (high or low), can represent a health risk, including the nonmetals fluorine and arsenic, the metals cadmium, lead, mercury, and zinc, and the radioactive elements uranium, potassium, thorium, and radon.

Eolian processes pertain to by the transport of sediment by wind. Eolian activity in Canada is significant and its long-term effects can threaten human health and

de diamètre, dont les bordures soulevées et renversées entourent une dépression en forme de cuvette, et les structures «complexes», dont le diamètre est généralement supérieur à 3 km, avec un soulèvement central en forme de pic ou d'anneau, un fossé annulaire et une bordure affaissée. Cependant, la morphologie n'est pas ce qui permet de confirmer l'origine des cratères d'impact. Ainsi, les petits cratères d'impact se reconnaissent à la présence de fragments de météorites et les grands cratères d'impact, à la présence d'une déformation caractéristique des roches appelée «métamorphisme d'impact». La formation de cratères d'impact de diamètre aussi petit que 20 km est susceptible d'éjecter de la poussière dans l'atmosphère qui réduit la luminosité du soleil et entraîne des perturbations de température affectant le climat et la biosphère à divers degrés à l'échelle de la planète. La chute de météorites dans les océans peut engendrer des tsunamis dont l'amplitude peut être considérablement supérieure aux tsunamis provoqués par les séismes. L'hypothèse selon laquelle l'impact d'un énorme météorite à la frontière du Crétacé et du Tertiaire serait à l'origine de l'extinction massive d'animaux survenue au cours de cette période est de plus en plus admise.

Le niveau de la mer le long des côtes canadiennes a oscillé sur des milliers d'années sous l'action principalement du glacio-eustatisme et de l'eustatisme. La présence simultanée de ces deux phénomènes produit des schémas de modifications du niveau de la mer qui varient dans l'espace et dans le temps le long des côtes. Dans les zones où il y a baisse du niveau de la mer, comme dans la baie d'Hudson, la ligne de rivage se déplace lentement vers le large de sorte que d'anciennes plages se trouvent actuellement à de nombreux mètres au-dessus du rivage actuel et à de nombreux kilomètres à l'intérieur des terres. Dans un grand nombre de régions, comme sur la côte est de la Nouvelle-Écosse, la montée du niveau de la mer rend le littoral vulnérable à l'érosion et à l'onde de tempête, particulièrement lors de tempêtes d'une violence exceptionnelle. L'érosion du littoral peut entraîner la perte de biens et d'infrastructures, et les inondations provoquées par les ondes de tempête sont susceptibles d'endommager les terres agricoles, les biens et les infrastructures. Le schéma actuel du changement du niveau de la mer le long des côtes canadiennes pourrait être sensiblement modifié au cours du prochain siècle en raison de la dilatation thermique des océans et de la fonte des calottes glaciaires du Groenland et de l'Antarctique, provoquées par les incidences liés au changement climatique.

La géochimie environnementale est l'étude de l'abondance, de la répartition et de la mobilité des éléments contenus dans le substratum rocheux et dans le gravier, le sable, le silt et les sédiments argileux qui forment les dépôts superficiels de la Terre. Un vaste éventail de caractéristiques géochimiques dans notre paysage sont indicateurs de variations dans la composition des matériaux superficiels et du substratum rocheux ainsi que de l'histoire géomorphologique. Dans le présent bulletin, la géochimie environnementale est examinée sous l'angle géologique et les thèmes associés aux risques sont présentés brièvement. L'accent est mis sur un petit groupe d'éléments, dont des éléments non métalliques (fluor et arsenic), métalliques (cadmium, plomb, mercure et zinc) et radioactifs (uranium, potassium, thorium et radon), qui, lorsque présents en concentrations anormales (élevées ou faibles), peuvent présenter des risques pour la santé.

Les processus éoliens sont liés au transport des sédiments par le vent. Au Canada, l'activité éolienne est considérable et ses incidences à long terme présentent une menace pour la santé humaine ainsi que damage agricultural and recreational lands. The hazards from eolian activity include wind erosion of agricultural soils, dust storms, sand-dune activity caused by human disturbance, overgrazing and/or drought, deflation areas, and alkali dust emissions. Human activities can contribute directly to eolian hazards by initiating or accelerating the rate of eolian processes. Although eolian processes are currently under control in most regions of Canada, wind erosion and sand-dune activity could increase in some regions because of increasing aridity associated with climate change.

About half of the Canadian landmass is covered by permafrost. In the far north, permafrost underlies the land continuously and can be up to 700 m thick. It thins gradually southwards and becomes increasingly discontinuous, until it eventually disappears. Permafrost is very sensitive to disturbances of the overlying ground surface because of changes to the ground thermal regime. Where ice-rich permafrost is present, forest fires or land clearance for example can cause thaw and ground settlement that lead to ground stability problems. Permafrost presents a challenge to development in the Canadian north, where different design and construction strategies are used in the construction of transportation systems, buildings and foundations, utilities, pipelines, and dams and dykes. Climate change has potential effects on permafrost.

Substantial amounts of gas hydrates are present in sediments within permafrost and offshore areas of Canada where suitable ground temperatures and pressure conditions exist. Known Canadian gas hydrate fields represent sizeable amounts of natural gas that form an important national energy reserve. The presence and stability of gas hydrates at a given location relate to past climates and geological history as well as to gas chemistry and existing temperature and pressure conditions. Gas hydrates may pose a hazard during exploration drilling or hydrocarbon production. The disassociation of gas hydrate in offshore areas can trigger large-scale submarine landslides and can be an important source of atmospheric greenhouse gas.

ACKNOWLEDGMENTS

This bulletin represents the major deliverable for project 950031, the National Geological Hazard Synthesis Project. Individual papers were reviewed by Gail Atkinson (Carleton University), Joe Desloges (University of Toronto), Ian Hutchison (Simon Fraser University) Michael Smith (Carleton University) and Richard Coles, Larry Dyke, Don Lemmen, Bernard Pelletier, Mark Pilkington and Glen Woodsworth (GSC). Pavel Kurfurst provided advice and savvy throughout all phases of this project. Ultimately, the success of this project directly reflects the efforts of the authors, all dedicated and very busy scientists, who contributed the papers.

pour les terres agricoles et les terres de loisirs. Les risques liés à l'activité éolienne sont, entre autres, l'érosion des terres agricoles, les tempêtes de poussière, l'édification et la destruction de dunes de sables engendrées par l'activité humaine, le surpâturage et la sécheresse, la formation de zones de déflation et les émissions de poussière alcaline. Les activités humaines peuvent contribuer directement aux risques liées à l'activité éolienne en faisant démarrer ou accélérer le rythme des processus éoliens. Bien que ces processus soient actuellement neutralisés dans la plupart des régions du Canada, l'érosion éolienne et l'édification et la destruction de dunes de sable sont susceptibles de s'accroître dans certaines régions en raison de l'accroissement de l'aridité associée au changement climatique.

La zone à pergélisol couvre environ la moitié de la masse continentale du Canada. Dans le Grand Nord canadien, le pergélisol est continu et peut atteindre une épaisseur de 700 m. Il s'amincit progressivement vers le sud et devient de plus en plus discontinu jusqu'à disparaître complètement. Il est très sensible aux perturbations que subit la surface sus-jacente du sol en raison des changements dans le régime thermique du sol. En présence d'un pergélisol riche en glace, les incendies de forêt ou le défrichement des terres peuvent engendrer le dégel et l'affaissement du sol, ce qui entraîne des problèmes de stabilité des sols. Le pergélisol présente un défi au développement du Nord canadien et l'on doit faire appel à diverses stratégies de conception et de construction pour la construction de réseaux de transport, d'édifices et de fondations, de services publics, de barrages et de digues. Le changement climatique est susceptible d'avoir des incidences sur le pergélisol.

Les hydrates de gaz sont présents en vastes quantités dans les sédiments des zones pergélisolées et extracôtières du Canada là où les conditions de température et de pression sont favorables. Les champs d'hydrates de gaz canadiens connus représentent des ressources appréciables de gaz naturel qui constituent d'importantes réserves énergétiques nationales. La présence et la stabilité des hydrates de gaz dans un lieu donné est liée aux climats du passé et à l'histoire géologique ainsi qu'à la chimie des gaz et aux conditions de température et de pression prévalant dans cette région. Les hydrates de gaz présentent un risque potentiel au cours de l'exécution de forages d'exploration ou de la production d'hydrocarbures. La dissociation des hydrates de gaz dans les régions extracôtières peut provoquer de vastes glissements de terrain sous-marins et l'émission importante de gaz à effet de serre dans l'atmosphère.

REMERCIEMENTS

Ce bulletin présente la pièce maîtresse du projet 950031, le Projet national de synthèse des risques géologiques. Les articles ont été révisés par Gail Atkinson (Université Carleton), Joe Desloges (Université de Toronto), Ian Hutchison (Université Simon Fraser), Michael Smith (Université Carleton) et Richard Coles, Larry Dyke, Don Lemmen, Bernard Pelletier, Mark Pilkington et Glen Woodsworth (CGC). Pavel Kurfurst a partagé son expérience et nous a guidé de ses conseils à toutes les étapes de ce projet. En bout de ligne, nous devons le succès du projet aux efforts de ces chercheurs dévoués qui, malgré leur emploi du temps très chargé, ont contribué à ces articles.

Seismicity and seismic hazards

John Adams[1] and Peter Basham[1,2]

Adams, J. and Basham, P., 2001: Seismicity and seismic hazards; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 7–25.*

Abstract: Many earthquakes occur along Canada's western margin near the edge of the North American Plate, but others occur within the plate in Arctic and southeastern Canada. Damaging earthquakes have occurred throughout history in both Eastern and Western Canada, and will undoubtedly recur in the future. Earthquakes have varied geological effects including ground rupture and land-level change, tsunamis, sediment liquefaction, and landslides; past earthquakes, including the prehistoric great Cascadia earthquake of AD 1700, have been studied by these effects. Historical earthquakes have caused damage to buildings and other structures, chiefly from direct shaking; future earthquakes may result in foundation failures and fires consequent on building damage. Damage can be mitigated by using earthquake-resistant designs. The Geological Survey of Canada has produced three generations of seismic hazard maps to assign hazard levels across the country; the fourth, which includes recent research developments, is nearing completion.

Résumé : De nombreux séismes se déclenchent le long de la marge occidentale du Canada, près de la frontière de la plaque nord-américaine, mais également à l'intérieur de la plaque dans l'Arctique et le sud-est canadien. Des séismes destructeurs se sont produits au cours de l'histoire dans l'Est et l'Ouest canadiens et se reproduiront sans aucun doute dans l'avenir. Les incidences géologiques des séismes sont variées, comprenant notamment la formation de fissures et de dénivellations dans le sol, le déclenchement de tsunamis, la liquéfaction des sédiments et les glissements de terrain. Ces incidences permettent aux sismologues d'étudier les séismes du passé comme, par exemple, l'important séisme historique qui s'est produit dans les Cascades en 1700 de notre ère. Les séismes historiques ont endommagé les constructions et d'autres structures en les secouant; les futurs tremblements de terre sont susceptibles de provoquer des accidents de fondations et des incendies liés aux dommages subis par les édifices. Ces dégâts peuvent être atténués par l'utilisation de structures parasismiques. La Commission géologique du Canada a dressé trois générations de cartes des risques sismiques afin d'attribuer des niveaux de risque dans les diverses régions du pays. Elle met le point final actuellement à la quatrième génération de carte de risques en tenant compte des derniers progrès de la recherche.

---

[1] Natural Earthquake Hazards Program, GSC, Pacific

[2] Provisional Technical Secretariat, Comprehensive Nuclear-Test-Ban Organization, Vienna Internaitonal Centre, P.O. Box 1200, A–1400 Vienna, Austria

INTRODUCTION

Background

The Earth's crust is made up of a dozen large and many smaller tectonic 'plates', which move slowly at different velocities and in different directions relative to one another. This movement creates boundaries where adjoining plates converge, diverge, or slide past each other. On a global scale, 98% of all earthquakes occur near plate boundaries because of these interactions. Few earthquakes occur at plate centres because these areas are located far from plate boundaries.

In Canada, the most seismically active areas are along the west coast where the Pacific Plate is sliding northward past the North American Plate, producing both the San Andreas Fault in California and the similar Queen Charlotte Fault off the west coast of Canada (Fig. 1). This situation is complicated off the coast of southwestern Canada and the northwestern United States by a small plate, the Juan de Fuca Plate, that is being pushed (or subducted) under North America, producing the Cascadia subduction zone (Fig. 1). Central Canada has very few earthquakes, but there is moderate seismicity along the eastern and northern continental margins and in southeastern Canada from eastern Ontario to New Brunswick.

Most of the largest earthquakes occur on plate boundaries and represent the release of strain accumulated as the plates continue to move, but lock together at their edges and suddenly slip. However, many earthquakes, including those in the plate interiors, represent the release of accumulated strains produced by internal deformation of the plate as it is stressed by interacting with all its neighbouring plates. Throughout eastern North America, this strain results from the North American Plate being pushed westward from the Mid-Atlantic Ridge, which forms the eastern boundary of the plate in a region of plate divergence (Fig. 1). Physically, each earthquake is the result of a sudden movement of two blocks of rock along a break (fault) deep within the Earth's crust. After very strong earthquakes, which involve movement on large faults, the fault displacement may be visible on the Earth's surface. It is rare to find such surface expression of earthquakes in Canada, as onshore ruptures on earthquake faults are generally too deep or too small to break the surface.

Earthquake intensity

Before the existence of instruments to record seismic waves, earthquakes were ranked in terms of the severity of the effects they caused. The early seismologists introduced the intensity scale as a qualitative measure of the effects of an earthquake on people, structures and natural features. The intensity scale most widely used today is the modified Mercalli scale, which

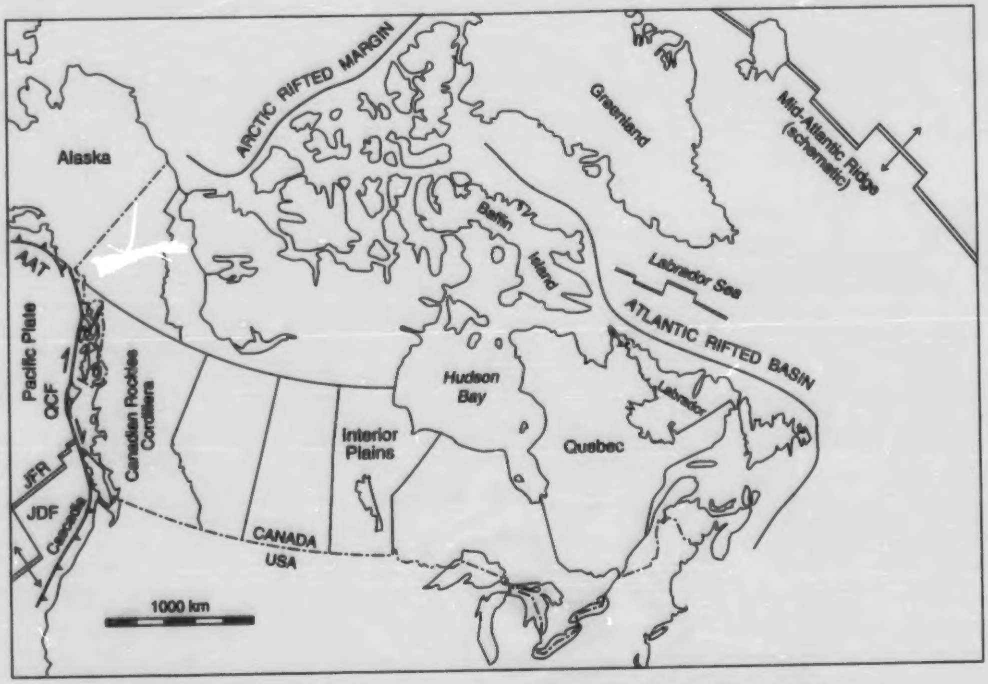


Figure 1. Map showing the tectonic plates that form the surface of the Earth's crust in Canada. AAT = Alaska-Aleutian subduction zone; QCF = Queen Charlotte Fault; JFR = Juan de Fuca spreading ridge; JDF = Juan de Fuca Plate. Arrows show direction of movement on plate boundaries.

has 12 levels of intensity, generally denoted by roman numerals, ranging from imperceptible shaking to catastrophic damage. Similar levels of intensity are contoured around an earthquake epicentre to produce an isoseismal map, i.e. a map showing contours of equal levels of damage or strength of shaking. A number of these are shown in Appendix A. The isoseismal maps for some of the older earthquakes are based on the slightly different Rossi-Forel intensity scale.

Earthquake magnitude

To overcome the limitations and subjectivity involved in assigning earthquake size on the basis of intensities, an instrumental scale of earthquake size was introduced by C.F. Richter in 1935. The instrumental magnitude scale is based on the principle that the amplitude of the recorded seismic waves reflects the strength of the earthquake at its source. Richter defined earthquake magnitude as the common logarithm of the maximum amplitude of the seismic waves recorded by a standard seismograph and corrected to a fixed reference distance to compensate for the decay of ground motion amplitudes with distance from the source. Richter's initial scale was derived for shocks recorded in southern California. Subsequently, other magnitude scales were defined for application to more distant earthquakes, and for earthquakes in regions where the efficiency of seismic-wave transmission differs from that in southern California. The most recent magnitude scale in common use is based on the seismic moment, which relates earthquake magnitude to the displacement and area of the fault rupture generating the earthquake.

Although the intent with each of the magnitude scales was to have a numerically equivalent measure of earthquake size, when the different scales are applied to the same earthquakes the results may differ by several tenths of a magnitude unit, because of differences in the relative excitation of the different types of seismic waves. Several different magnitude scales are used by the Geological Survey of Canada to characterize Canadian earthquakes, but we will not burden the reader with further distinctions in this paper. We will simply refer to 'magnitude' or 'M'.

History of earthquake recording in Canada

The first permanent seismograph stations in Canada capable of recording large earthquakes were established in 1897 in Toronto, in 1898 in Victoria, and in 1906 in Ottawa. It was not until after the strong 1925 St. Lawrence River valley and 1946 Vancouver Island earthquakes that more sensitive seismographs suitable for recording nearby earthquakes were set up in Eastern and Western Canada. The first national seismograph network was established in the 1960s and expanded considerably in the 1970s. The Geological Survey of Canada (GSC) now operates 90 seismograph stations throughout Canada (Fig. 2). The data from most of these stations is

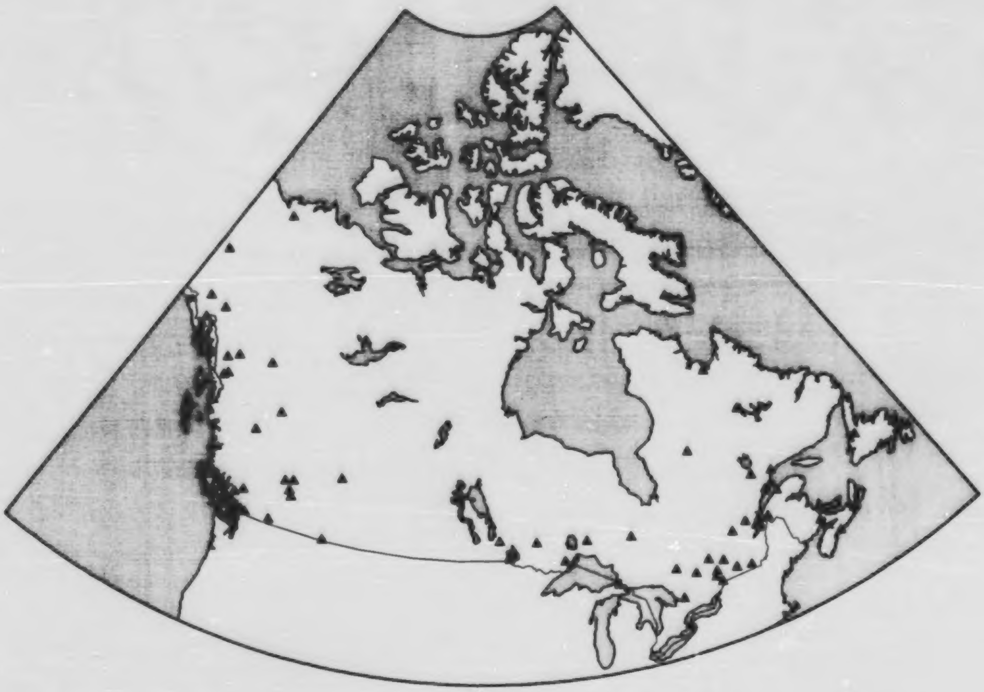


Figure 2. Canadian seismograph network operated by the Geological Survey of Canada in 1997.

transmitted to GSC analysis centres in Ottawa, Ontario, and/or Sidney, British Columbia, allowing rapid analysis and reporting of significant earthquakes as they occur.

Information on earthquakes that occurred prior to the establishment of appropriate seismographs has been obtained from written accounts (diaries, newspapers, etc.) of earthquake shaking and damage. This historical record extends back to the seventeenth century in Eastern Canada, but only to the early part of the nineteenth century in Western Canada. For the earliest historical periods, because of sparse population and limited reporting, only the largest earthquakes have been documented. The approximate epicentral locations and magnitudes of older earthquakes have been determined from isoseismal information, for example by establishing relationships between the areas inside isoseismal contours and magnitudes for modern earthquakes and applying this information to establish the approximate magnitudes of the historical earthquakes.

Seismicity map of Canada

The earthquakes plotted on the accompanying seismicity map of Canada (Fig. 3, in pocket) are from the Seismic Hazard Earthquake Epicentre File, which contains information on 39 000 earthquakes from 1568 to 1997. To produce a fair representation of seismicity in Canada, earthquakes were selected by date to account for our improvements in detecting earthquakes across Canada, and screening was used to remove some of the many overlapping symbols. A smaller version of this map is provided in Figure 4 for easy reference. The area covered extends beyond the Canadian border (e.g. into Alaska, Washington, and New England) to include earthquake zones that could affect Canadian territory. Information on some of the larger and damaging Canadian earthquakes is given in Figure 3 and information on specific damaging earthquakes is provided in Appendix A.

REGIONAL UNDERSTANDING OF EARTHQUAKE ACTIVITY

Western Canada

The most intense earthquake activity occurs along the coast of British Columbia and is caused by the interaction of the Pacific, North American, and Juan de Fuca tectonic plates. In the Vancouver Island–lower mainland region, the seismicity

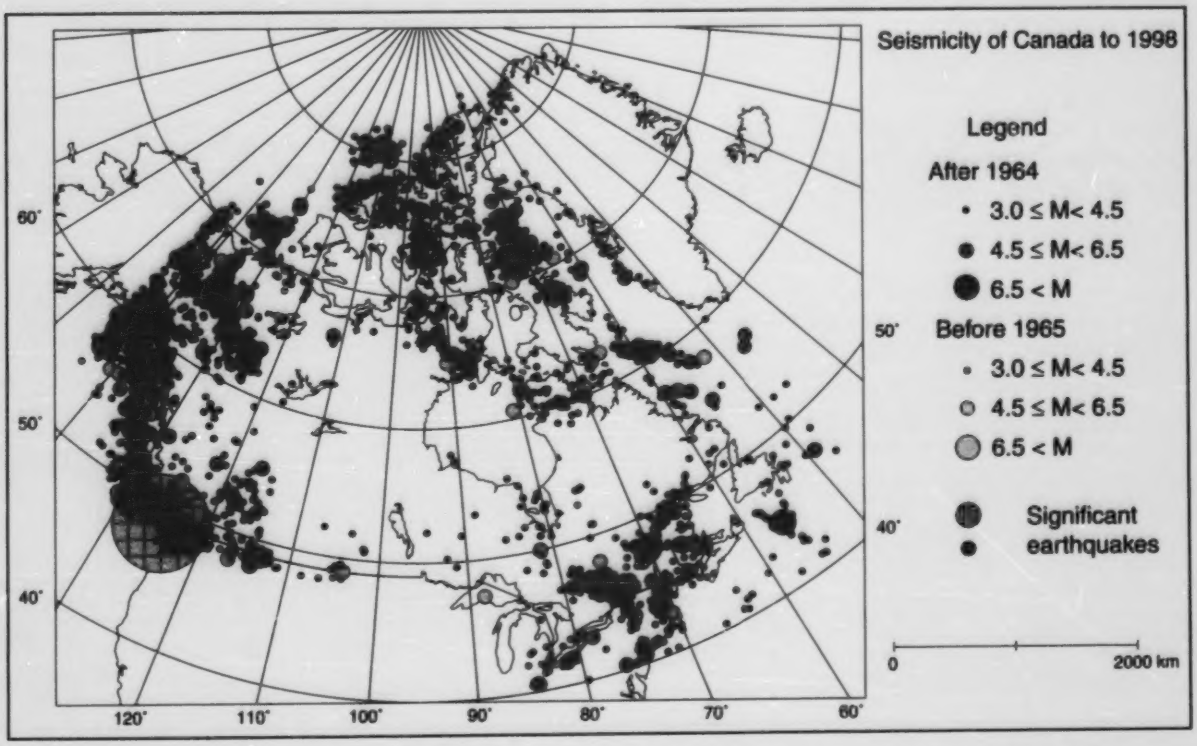


Figure 4. Grey-scale version of the 1998 seismicity map of Canada. The map is produced using a selection of earthquakes from the Geological Survey of Canada's Seismic Hazard Earthquake Epicentre File, which includes pending revisions to the Canadian Earthquake Epicentre File (CEEF) and selected earthquake epicentres in the United States. The Canadian Earthquake Epicentre File contains information on more than 39 000 earthquakes known to have occurred in or near Canada from 1568 to 1996.

is associated with the Cascadia subduction zone (*see* 'Background', above), where earthquakes such as the one that occurred in 1909 (M6.0; Fig. 3) occur within the subducting Juan de Fuca Plate. The 1946 earthquake (M7.3; Fig. 3 and Appendix A), in the overriding North American Plate, is the largest that occurred in the twentieth century in southwestern British Columbia. Geological evidence on Vancouver Island and evidence from a tsunami recorded in Japan indicate that a prehistoric great subduction earthquake occurred at about 21:00 h on 26 January 1700 immediately offshore of Vancouver Island (M9; Fig. 3). Geological evidence of this and prior earthquakes indicates that such great earthquakes happen about every 500 to 600 years on average, although repeat times between events vary. The Queen Charlotte Fault off the west coast of the Queen Charlotte Islands has been the site of several major earthquakes, including Canada's largest historical earthquake in 1949 (M8.1; Fig. 3 and Appendix A).

Farther north in the St. Elias region of the Yukon Territory and southeastern Alaska, motion along the plate boundary undergoes a transition from strike-slip (i.e. the two plates are sliding past one another) to convergence (i.e. involving subduction). An unprecedented sequence of three major earthquakes occurred in the Yakutat Bay area in 1899 (M7.9; Fig. 3). The plate interaction in this region results in a significant number of earthquakes in the southwestern Yukon Territory and adjacent areas of northwestern British Columbia.

Seismicity occurs inland across the Cordillera except in north-central British Columbia, where very few earthquakes occur. Seismicity is more intense in the northern Cordillera, particularly in pockets in the Richardson Mountains, northern Yukon Territory, and in the Mackenzie Mountains, western Northwest Territories. The largest known earthquake in the eastern Cordillera occurred in the Nahanni region of the Northwest Territories in 1985 (M6.9; Fig. 3 and Appendix A).

Southeastern Canada

To a first approximation, the stress field in Canada east of the Cordillera is compressional from the northeast, due to ocean-floor spreading on the Mid-Atlantic Ridge at the eastern margin of the North American Plate (Fig. 1). This stress field causes thrust or strike-slip earthquakes along pre-existing faults in the upper 30 km of the crust. (Seismologists consider earthquakes representing shortening of the Earth's crust to be thrust earthquakes; these earthquakes occur on reverse faults where the block forming the upper surface of the fault plane moves upward relative to the underlying rock. To a geologist, a thrust fault is a reverse fault that generally dips at less than 45°; seismologists do not make this distinction. Furthermore, many thrust earthquakes in Eastern Canada involve shortening motion on ancient normal faults and so reverse the sense of motion that created the fault in the first instance.) Most of the significant earthquakes on land can be associated with the ancient rifted edge of the North American continent. This edge currently lies along the St. Lawrence River valley from Montréal to Sept-Îles; related rift structures lie along the Ottawa and Saguenay river valleys. The seismicity nearly always involves thrust earthquakes at depths of 5 to 30 km within the Precambrian shield and occurs mainly in three clusters: western Quebec, which includes the 1732 (M5.8), 1935 (M6.2), and 1944 (M5.8) earthquakes (*see* Fig. 3 and Appendix A); Charlevoix, northeast of the city of Quebec, a source of numerous large earthquakes (1663, 1791, 1860, 1870, and 1925; M6.0 to 7.0; *see* Fig. 3 and Appendix A) and a continuous source of smaller events; and the Lower St. Lawrence region near Baie-Comeau, a diffuse cluster of mostly small earthquakes. Moderate activity occurs in the Appalachian region throughout most of New Brunswick. The most recent widely felt earthquake in Eastern Canada occurred in the Saguenay region in 1988 (M5.9; Fig. 3 and Appendix A).

The eastern margin

Along the eastern continental margin, seismicity includes the 1929 Grand Banks (M7.2; Fig. 3 and Appendix A) and the 1933 Baffin Bay (M7.3; Fig. 3 and Appendix A) earthquakes. These and smaller events are concentrated at the ocean-continent transition zone and appear to have been caused by the reactivation of old rift faults formed during the opening of the Atlantic Ocean and Baffin Bay tens of millions of years ago. In the middle part of the Labrador Sea, earthquakes are associated with an extinct spreading ridge and its associated transform faults (i.e. an ancient equivalent of the present-day Mid-Atlantic Ridge).

Arctic Canada

In this region, earthquakes occur mainly on Baffin Island, along an arcuate band between the Boothia (Nunavut) and Ungava (northern Quebec) peninsulas (including the 1989 Ungava earthquake), in the Sverdrup Basin under the central Arctic Islands, and in the Beaufort Sea. The 1989 Ungava earthquake in northern Quebec (M6.3; Fig. 3 and Appendix A) ruptured the surface and created a 10 km long fault scarp up to 2 m high, the first surface rupture associated with a historical earthquake in eastern North America. The Baffin and Boothia–Ungava earthquakes are spatially associated with steep gradients in the rate of postglacial uplift, which suggests that they may occur because of differential glacio-isostatic uplift. The Baffin Island earthquakes are unique in Eastern Canada in that some of them involve normal (i.e. extensional) faulting. The Sverdrup Basin earthquakes represent strike-slip deformation beneath a thick accumulation of sedimentary rocks. The passive Arctic Ocean margin has a rifted ocean-continent transition comparable to the Atlantic margin, but it seems to be seismically active only where it has been recently loaded by thick, geologically young sediments, most notably in the Beaufort Sea. Most information on Arctic earthquakes dates from the completion of the national seismograph network in 1965 to the mid 1990s when key stations at Alert and Mould Bay were closed; it is quite likely that not all active zones have been identified.

NEOTECTONICS AND PALEOSEISMOLOGY

Neotectonics, the study of recent tectonic processes, provides important insight into earthquake occurrence and cause. The Canadian landmass includes active convergent and strike-slip plate boundaries, several major mountain systems, two passive continental margins, and a stable craton. Correlations between tectonics and physiography are strongest in the west where the landscape bears a strong imprint of convergent and strike-slip plate regimes and earthquakes are abundant (Fig. 1, 3). By contrast, the neotectonic imprint in Eastern and northern Canada is enigmatic, and although major concentrations of earthquakes in many areas are associated with reactivated, early Phanerozoic (up to 570 million years old) structures, only limited late Quaternary (younger than 12 000–20 000 years) faulting has been discovered. Despite its very low seismicity, the vast Canadian craton is deforming (chiefly uplifting) glacio-isostatically at a moderate rate as a result of melting of the Laurentide Ice Sheet thousands of years ago.

Paleoseismology

Paleoseismology, a relatively new interdisciplinary field of Earth science research, is essentially the study of prehistoric earthquakes through the interpretation of their effects preserved in the geological record. Although the evidence is often sparse and hard to interpret with certainty, this research can pay large dividends for seismic hazard assessment.

The most important contribution of the past decade has been the accumulation of geological evidence for the last great Cascadia subduction earthquake under Vancouver Island. This evidence involves suddenly submerged coastal marshes covered by clean sand layers, interpreted as representing land subsidence during the earthquake followed by a tsunami. Radiocarbon dating of marsh peat placed the last earthquake about 300 years ago, and the resulting tsunami was then identified from Japanese harbour records. The earthquake occurred at about 21:00 h on 26 January 1700, which is consistent with native oral histories of a devastating earthquake one winter night. Other geological evidence, from the turbidite deposits of extensive submarine landslides triggered by the earthquake, indicates that such earthquakes happen about every 500–600 years on average, although the time interval between events varies. Establishing a chronology of past events should enable a reliable estimate of the chances of the next great earthquake occurring in the near future.

The accumulation of geological evidence in southwestern British Columbia is leading to the identification of past large crustal earthquakes and may allow the identification of some of the many active faults, so allowing a better assessment of the seismic hazard than is possible from the short historical earthquake record alone. In Eastern Canada, a few paleoearthquakes have been deduced from lake-sediment studies and other geological evidence of the shaking they have caused.

Earthquakes and faults

Earthquakes occur on faults, usually deep underground. Geologists in California, New Zealand, and many other places have mapped the surface trace of some of the larger faults, and seismologists talk about this or that earthquake happening 'on the San Andreas Fault'. Hence, there is a widespread misconception that every earthquake can be associated with a known fault. The situation onshore in Canada is less simple. Our earthquakes happen far from plate boundaries and on faults that have moved little and seldom, unlike California where faults like the San Andreas have moved many kilometres over the millennia and are clearly seen on the Earth's surface. Although Canada's crust is disrupted by many geological faults, most of these are inactive, and we are as yet unable to identify the ones that will move in the near future. Thus, realistic estimates of earthquake occurrence should be made using probabilistic (statistical) methods rather than deterministic ones, which depend on knowledge of the causative faults.

SEISMIC HAZARDS

Effects of earthquakes and types of seismic hazard

Small earthquakes (M3 and less) are often not even felt; moderate earthquakes (M3–4.5) are usually felt and may cause minor damage; larger earthquakes may cause significant to great damage, depending on size and distance from the epicentre and on the characteristics of the ground and building construction. The effects of damaging earthquakes fall into several classes.

Ground rupture

Some earthquakes break and displace the surface of the Earth along faults, tearing apart buildings built across them, rupturing underground utilities, and changing the underground hydrogeology. Fortunately, such surface faulting is extremely rare in Canada; the only confirmed example on land is from the 1989 Ungava earthquake, which occurred in an unpopulated area.

Land level change

Great earthquakes deform the surface of the land over large areas. In 1700, the most recent great earthquake on the Cascadia subduction zone caused the west coast of Vancouver Island to subside about 1 m, inundating coastal marshes and forests with sea water.

Tsunamis

Great earthquakes that cause massive land level change underwater also generate large tsunamis that strike adjacent coasts. These waves also propagate across the ocean to damage its far shores. The 1700 Cascadia earthquake caused a large, local tsunami on the west coast of Vancouver Island. Distant earthquakes in 1960 (Chile) and 1964 (Alaska) have also caused tsunami damage in British Columbia. Large earthquakes may trigger large submarine landslides, which also generate damaging tsunamis; the 1929 tsunami that struck southern Newfoundland was a recent example. More information on tsunamis is given in Clague (2001).

Ground shaking

Earthquakes represent the sudden release of energy along an underground fault in the Earth's crust. This energy is propagated through the Earth as seismic waves that create vibrations on the surface that are felt as shaking. In general, the strength of shaking on a rock site decreases with distance, so the farther away from a given earthquake, the weaker the shaking. Earthquake shaking is the primary cause of building failure, which in turn causes loss of life.

Amplification of shaking on soft ground

When earthquake waves pass from crustal rock to overlying sediment and soil at the surface, they slow down and their amplitude increases. Thus a house on deep soil in a river valley may be shaken much more strongly than a nearby house on a rock foundation. The extra degree of shaking depends on the type and thickness of the soil, but can approach a five-fold increase.

Liquefaction

Certain types of saturated soil lose their strength and liquefy during earthquake shaking. A sediment-water mixture may be ejected from the ground, forming small-scale sand 'volcanoes' or 'dykes'. Building foundations constructed on such soils may subside, so damaging the structure.

Consequent fires

Earthquakes cause damage not only from ground shaking, but also from fires that are started accidentally through upset appliances and broken gas mains. A loss of communications, broken water lines, and limited access through rubble-choked streets can often allow such fires to spread uncontrollably. Losses due to ground shaking in the 1995 Kobe earthquake were greatly compounded by fire damage.

Landslides

Strong shaking of hilly terrain may trigger numerous landslides. These may damage structures directly, or disrupt communication routes such as railways and roads along valley floors. Large landslides in river valleys may dam the river, creating a landslide-dammed lake. Lakes formed by large slides across small rivers may be stable, permanent features, but larger rivers may quickly overtop and wash away the debris dam, releasing a catastrophic flood. More information on landslides is given in Evans (2001).

History of damage from Canadian earthquakes

Many of the significant Canadian earthquakes have produced minor damage (*see* Figure 3). Total losses have not been high to date because of the low population density in most of the country. The tsunami from the 1929 Grand Banks earthquake caused the highest number of deaths (27) from a single earthquake; the more recent 1988 Saguenay earthquake caused about $20 million in damage in Quebec. Appendix A summarizes and describes the more significant losses. An earthquake disaster has yet to occur in Canada, although the past history of significant damage and statistical inferences from smaller earthquakes indicate that chances are uncomfortably high that a major earthquake will occur close to a Canadian urban centre sometime in the next 50 years, most likely in southwestern British Columbia or the Ottawa River–St. Lawrence River valleys.

Seismic hazard maps

Although large earthquakes cannot be prevented, their effects can be mitigated by constructing buildings so that they survive earthquakes expected to occur. To mitigate future disasters, buildings in seismic regions of the country must be built to an earthquake-resistant design specified by the National Building Code of Canada. The earthquake provisions of the Code provide a formulation for the seismic forces to which buildings must be designed, based on the seismic zones depicted on the two zoning maps shown in Figure 5. These zoning maps are prepared by the Geological Survey of Canada from a consideration of the seismic history of the country and take into account the state of knowledge about tectonics and the geological controls on earthquake activity. The maps present ground shaking with a 10% probability of being exceeded in 50 years (0.0021/a). The map of peak ground acceleration is a high-frequency measure relevant to short, stiff buildings and the map of peak ground velocity is a low-frequency measure relevant to tall, flexible buildings.

Such seismic zoning maps are updated every 10 to 15 years to include improvements in our understanding of the seismicity and seismotectonics and in the structural aspects of earthquake-resistant design. The current seismic zoning maps shown in Figure 5 may be replaced in the 2003 edition of the National Building Code by the fourth-generation maps derived by the GSC, which are currently under review by the Canadian engineering community. This new formulation of seismic hazards in Canada has been made available for review at our Web site (*see* below).

ADDITIONAL INFORMATION

Additional information on seismicity, seismotectonics, and seismic hazards in Canada is available on the World Wide Web at http://WWW.seismo.nrcan.gc.ca and http://WWW.pgc.nrcan.gc.ca or from the following offices of the Geological Survey of Canada: National Earthquake Hazards Program, 7 Observatory Crescent, Ottawa, Ontario K1A 0Y3 and Pacific Geoscience Centre, P.O. Box 6000, Sidney, British Columbia V8L 4B2.

ACKNOWLEDGMENTS

This contribution builds upon the text that accompanies the 1989 and 1994 seismicity maps of Canada and material prepared for our Web pages. We thank F.M. Anglin and S. Halchuk for database preparation, M. Lamontagne and

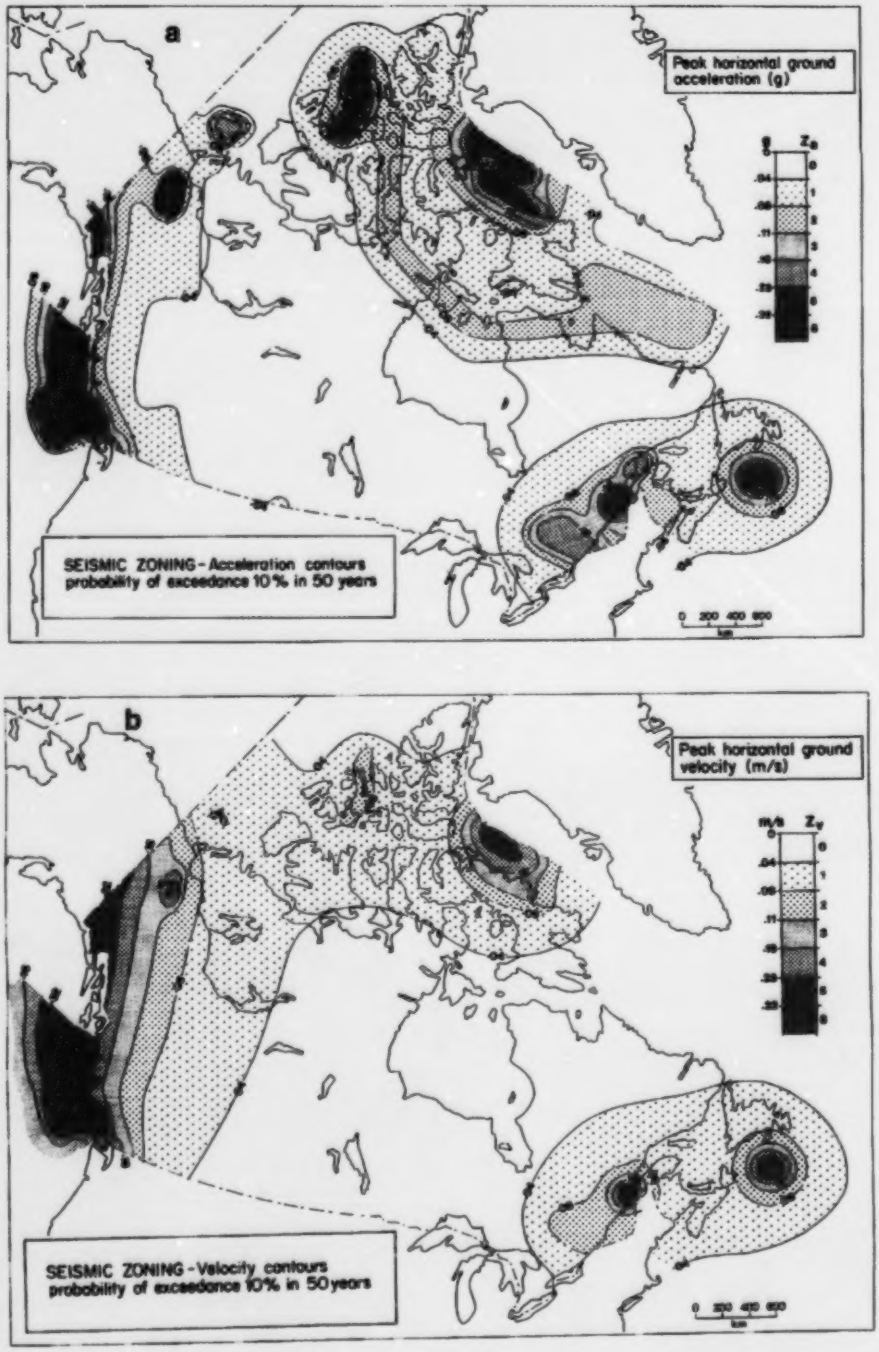

a
Peak horizontal ground acceleration (g)
g
Zₐ
0
.04
.08
.11
.16
.23
.32
0
1
2
3
4
5
6
SEISMIC ZONING - Acceleration contours probability of exceedance 10% in 50 years
0 200 400 600
km
b
Peak horizontal ground velocity (m/s)
m/s
Zᵥ
0
.04
.08
.11
.16
.23
.32
0
1
2
3
4
5
6
SEISMIC ZONING - Velocity contours probability of exceedance 10% in 50 years
0 200 400 600
km

Figure 5. Seismic zoning maps in the 1985 edition of the National Building Code of Canada. ***a****) Peak ground acceleration and* ***b****) peak ground velocity measures of ground shaking with a 10% probability of being exceeded in 50 years.*

G. Rogers for compiling the damage photographs, and R.G. North for co-ordinating the two previous editions of the Seismicity Map of Canada upon which our large-scale seismicity map is based.

REFERENCES

Anglin, F.M., Wetmiller, R.J., Horner, R.B., Rogers, G.C., and Drysdale, J.A.
1990: Seismicity map of Canada; Map 15 *in* Canadian Geophysical Atlas; Geological Survey of Canada, scale 1:10 000 000.

Clague, J.J.
2001: Tsunamis; *in* A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548.

Evans, S.G.
2001: Landslides; *in* A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548.

North, R.G.
1994: Canada — Seismicity; The National Atlas of Canada (fifth edition), Map MCR 4171, scale 1:7 500 000.

Smith, W.E.T.
1962: Earthquakes of Canada and adjacent areas 1534–1927; Publications of the Dominion Observatory, Department of Mines and Technical Surveys, Ottawa, v. 26(5), p. 271–301 (reprinted with amendments 1972).
1966: Earthquakes of Canada and adjacent areas 1928–1959; Publications of the Dominion Observatory, Department of Mines and Technical Surveys, Ottawa, v. 32(3), p. 87–121.

SUGGESTED READING

Adams, J.
1996: Paleoseismology studies in Canada - a dozen years of progress; Journal of Geophysical Research, v. 101, p. 6193–6207.

Adams, J. and Basham P.W.
1989: The seismicity and seismotectonics of Canada east of the Cordillera; Geoscience Canada, v. 16, p. 3–16.

Adams, J. and Bell, J.S.
1991: Crustal stresses in Canada; Chapter 20 *in* Neotectonics of North America, (ed.) D.B. Slemmons, E.R. Engdahl, M. Zoback, and D. Blackwell; Geological Society of America, The Geology of North America, v. DMV-1, p. 367–386.

Adams, J. and Clague, J.J.
1993: Neotectonics and large-scale geomorphology of Canada; Progress in Physical Geography, v. 17, p. 210–226.

Adams, J., Wetmiller, R.J., Hasegawa, H.S., and Drysdale, J.
1991: The first surface faulting from a historical intraplate earthquake in North America. Nature, v. 352, p. 617–619.

Basham, P.W.
1995: Recent advances in understanding of earthquake potential and seismic hazards in Canada; *in* Proceedings, Seventh Canadian Conference on Earthquake Engineering, 5–7 June 1995, Montréal, Quebec, p. 45–64.

Basham, P.W., Weichert, D.H., Anglin, F.M., and Berry, M.J.
1985: New probabilistic strong seismic ground motion maps of Canada; Bulletin of the Seismological Society of America, v. 75, p. 562–595.

Bruneau, M. and Lamontagne, M.
1994: Damage from 20th century earthquakes in eastern Canada and seismic vulnerability of unreinforced masonry buildings; Canadian Journal of Civil Engineering, v. 21, no. 4, p. 643–662.

Clague, J.J. (comp.)
1996: Paleoseismology and seismic hazards, southwestern British Columbia; Geological Survey of Canada, Bulletin 494, 88 p.

Hodgson, E.A.
1945: Industrial earthquake hazards in eastern Canada; Seismological Society of America, Bulletin, v. 35, no. 4, p. 151–174.

Hyndman, R.D.
1995: Giant earthquakes of the Pacific Northwest; Scientific American, v. 273, no. 6, p. 50–57.

Hyndman, R.D., Rogers, G.C., Dragert, H., Wang, K., Clague, J.J., Adams, J., and Bobrowsky, P.T.
1996: Giant earthquakes beneath Canada's west coast; Geoscience Canada, v. 23, p. 63–72.

Lamontagne, M.
1991: Les tremblements de terre au Québec; Association professionnelle des géologues et géophysiciens du Québec, Collection Environnement et Géologie, vol. 12, p. 265–289.

Lamontagne, M. and Bruneau, M.
1993: Impact of eastern Canadian earthquakes of 1925, 1929, 1935, and 1944; Earthquake Engineering Research Institute (EERI), slide set, 27 slides, 35 mm.

Rogers, G.C.
1994: Earthquakes in the Vancouver area; Geological Survey of Canada, Bulletin 481, p. 221–229.

Rogers, G.C. and Horner, R.B.
1991: An overview of western Canadian seismicity; Chapter 5 *in* Neotectonics of North America, (ed.) D.B. Slemmons, E.R. Engdahl, M. Zoback, and D. Blackwell; Geological Society of America, The Geology of North America, v. DMV-1, p. 69–76.

Stevens, AE
1980: History of some Canadian and adjacent American seismograph stations; Seismological Society of America, Bulletin, v. 70, p. 1381–1393.
1988: Earthquake hazard and risk in Canada; *in* Natural and Man-Made Hazards, (ed.) M.J. El-Sab. and H.S. Marty; D. Seidel Publishing Company, Dordrecht, Netherlands, p. 43–61.
1988: Perils séismiques au Canada; *in* Risk Assessment and Management: Emergency Planning Perspectives, (ed.) L.R.G. Martin and G. Lafond; University of Waterloo Press, Waterloo, Ontario, p. 21–41.
1991: Damage associated with three early eastern North American earthquakes; *in* Proceedings, Earthquake Engineering — Sixth Canadian Conference, Toronto, 12–14 June 1991; University of Toronto Press, Toronto, Ontario, p. 807–814.
1994: Earthquakes in the Lake Ontario region; Geoscience Canada, v. 21, p. 105–111.

APPENDIX A

Examples of isoseismal maps and damage from Canadian earthquakes

Date (y/m/d)	Latitude (°N)	Longitude (°W)	Magnitude	Comment
1732/09/16	45.5	73.6	5.8	Near Montréal, Quebec. Widely felt. The town walls were damaged slightly and a few large stone buildings needed significant repairs. Some accounts suggest that about 300 houses (probably about one half of those in Montréal at the time) were damaged. However, a recent analysis argues for less damage and that the economic impact of this earthquake was considerably less than that from three large fires in the preceding 40 years. No deaths or serious injuries have been confirmed
1925/03/01	47.8	69.8	6.2	Charlevoix–Kamouraska region, Quebec. Felt over most of Quebec, the Maritimes, southern Ontario, and parts of northeastern United States (Fig. A-1). Considerable damage in the epicentral region, on both shores of the St. Lawrence River, numerous chimneys were broken on buildings sited on deep alluvial soils. In addition, in Saint-Urbain and Rivière-Ouelle (Fig. A-2), churches were severely damaged. In the city of Quebec, buildings were damaged in the lower town and in Shawinigan, some 250 km from the epicentre, houses and a church located on thick clay deposits were damaged due to seismic wave amplification.
1929/11/18	44.5	56.3	7.2	Atlantic Ocean south of Newfoundland. Felt over a wide area of eastern North America (Fig. A-3), the earthquake caused a large underwater landslide that broke 12 transatlantic telegraph cables and generated a tsunami. Twenty-seven people were drowned and much damage (Fig. A-4) was caused by the 5 m high tsunami on Burin Peninsula. Most serious loss of life in any recorded Canadian earthquake. On land, damage due to earthquake vibrations was limited to Cape Breton Island where chimneys were overthrown or cracked and where some highways were blocked by minor landslides.
1935/11/01	46.8	79.1	6.2	Quebec–Ontario border. Felt over much of Eastern Canada (Fig. A-5). Minor damage (Fig. A-6) at Témiscaming, Quebec, and North Bay and Mattawa, Ontario. Some 300 km away from the epicentre, near Parent, Quebec, earthquake vibrations triggered a 30 m slide of a railway embankment.
1944/09/05	45.0	74.8	5.8	Eastern Ontario–New York border. Widely felt (Fig. A-7). Although this earthquake was only of moderate magnitude, it caused damage totalling $2 million (1944 dollars) in Cornwall (Ontario), Massena (New York), and nearby communities. About 2000 chimneys and many brick homes were damaged in Cornwall (Fig. A-8). Considerable damage was caused at the Collegiate and Vocational School where falling masonry broke through the roof of the adjacent gymnasium.
1946/06/23	49.8	125.3	7.3	Central Vancouver Island. Widely felt (Fig. A-9). Extensive property damage (Fig. A-10, A-11) along the east coast of Vancouver Island; one person was drowned.
1949/08/22	53.6	133.3	8.1	Off the Queen Charlotte Islands. Canada's largest earthquake (M8.1) and one of the world's largest strike-slip earthquakes. Felt over a wide area of western North America (Fig. A-12). Some damage (Fig. A-13) on the Queen Charlotte Islands.
1985/12/23	62.2	124.2	6.9	Nahanni region, Northwest Territories. Widely felt in the Northwest Territories, Alberta and British Columbia. A smaller event (M6.6) that occurred in the same area two months earlier (October 5) triggered an immense rock avalanche (Fig. A-14) containing 5 to 7 x 10^6 m^3 of rock.
1988/11/25	48.1	71.2	5.9	Saguenay region, Quebec. The Saguenay earthquake was felt with maximum intensity Modified Mercalli (MM) VIII, but more typically MM VII, in the Chicoutimi–Jonquière–La Baie area, was felt strongly by most people within 500 km, was felt by many within 1000 km, and was perceptible by some people, under special circumstances, beyond 1000 km (Fig. A-15). In the sparsely populated epicentral area damage was modest, limited to cracked or fallen unreinforced masonry walls and a few minor landslides. Damage outside the epicentral area was correlated with underlying unconsolidated deposits. Damage at Jonquière, Chicoutimi, La Baie, Québec, and as far away as Montréal (Fig. A-16) totalled a few tens of million dollars.
1989/12/25	60.1	73.6	6.3	Ungava Peninsula, Quebec. The 1989 Ungava earthquake occurred in an uninhabited area of the Ungava Peninsula. Due to its shallow focus, the earthquake caused a surface rupture (Fig. A-17), the first earthquake in eastern North America to have produced surface faulting.

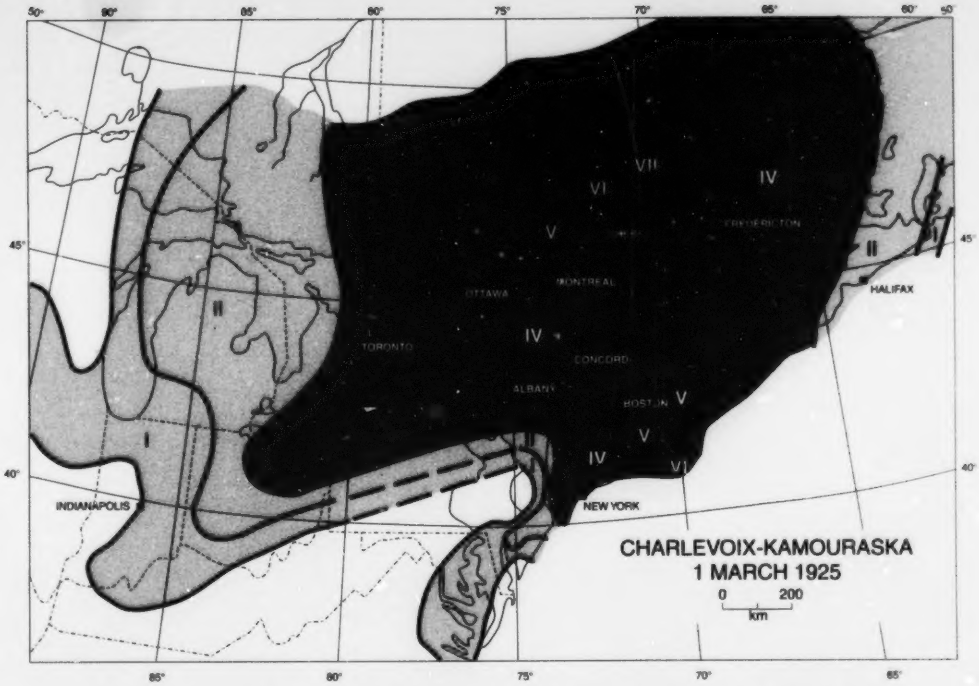

CHARLEVOIX-KAMOURASKA
1 MARCH 1925
0
200
km
VII
VI
IV
V
FREDERICTON
OTTAWA
MONTREAL
TORONTO
IV
CONCORD
ALBANY
BOSTON
V
V
IV
VI
II
HALIFAX
II
I
INDIANAPOLIS
NEW YORK
45°
40°
85°
80°
75°
70°
65°

Figure A-1. Isoseismal map (using the Rossi-Forel scale) for the 1925 Charlevoix-Kamouraska earthquake (from *Smith, 1962). Maximum intensity on the Rossi-Forel scale was IX ("partial or total destruction of some buildings").*

Figure A-2.

The 1925 Charlevoix–Kamouraska earthquake caused considerable damage to a church located on thick clay deposits at Rivière-Ouelle on the south shore of the St. Lawrence River. Photograph from *the archives of the Earth Physics Branch.*

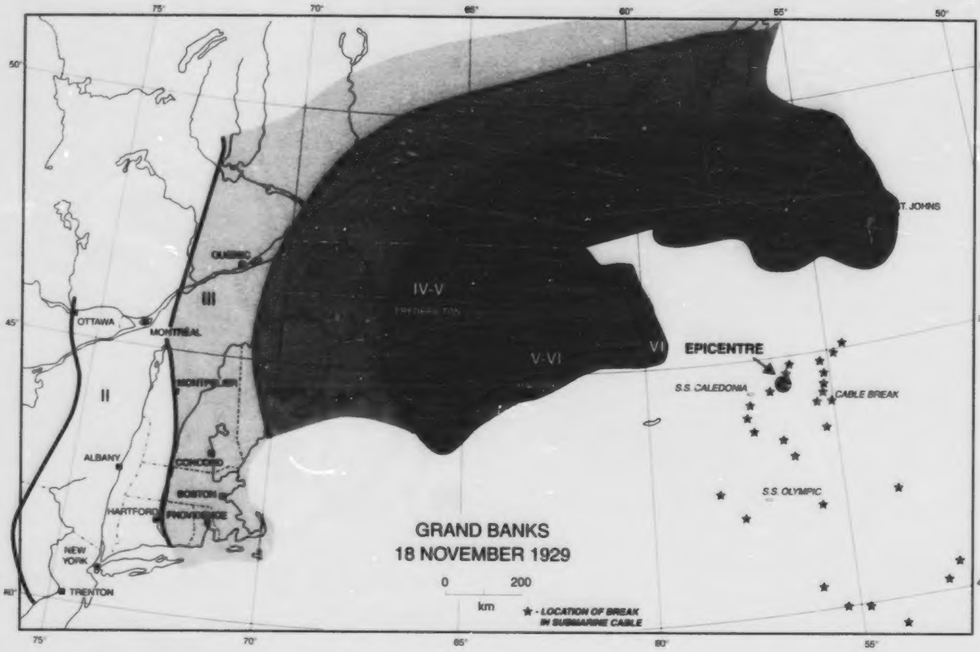

75°
70°
65°
60°
55°
50°
50°
45°
40°
OTTAWA
MONTREAL
QUEBEC
MONTPELIER
ALBANY
CONCORD
BOSTON
HARTFORD
PROVIDENCE
NEW YORK
TRENTON
ST. JOHNS
FREDERICTON
II
III
IV-V
V-VI
VI
EPICENTRE
S.S. CALEDONIA
CABLE BREAK
S.S. OLYMPIC
GRAND BANKS
18 NOVEMBER 1929
0
200
km
★ - LOCATION OF BREAK IN SUBMARINE CABLE

Figure A-3. *Isoseismal map for the 1929 Grand Banks earthquake* (from *Smith, 1966*). *All intensities are on the Rossi-Forel scale.*

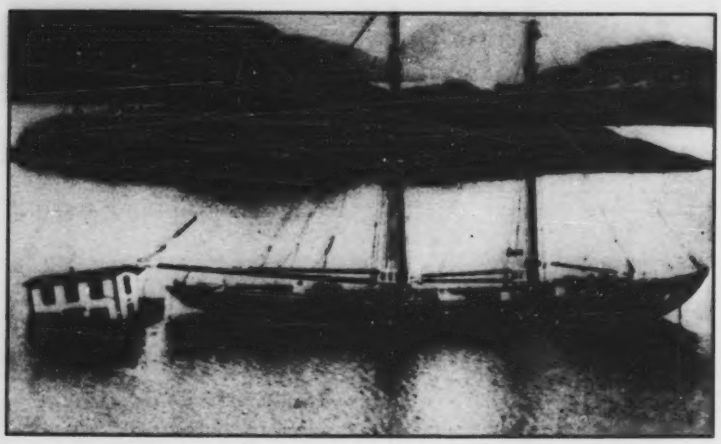


Figure A-4.

The tsunami from the 1929 Grand Banks earthquake washed this two-storey house 3 km out to sea in southern Newfoundland. Photograph from the archives of the Earth Physics Branch.

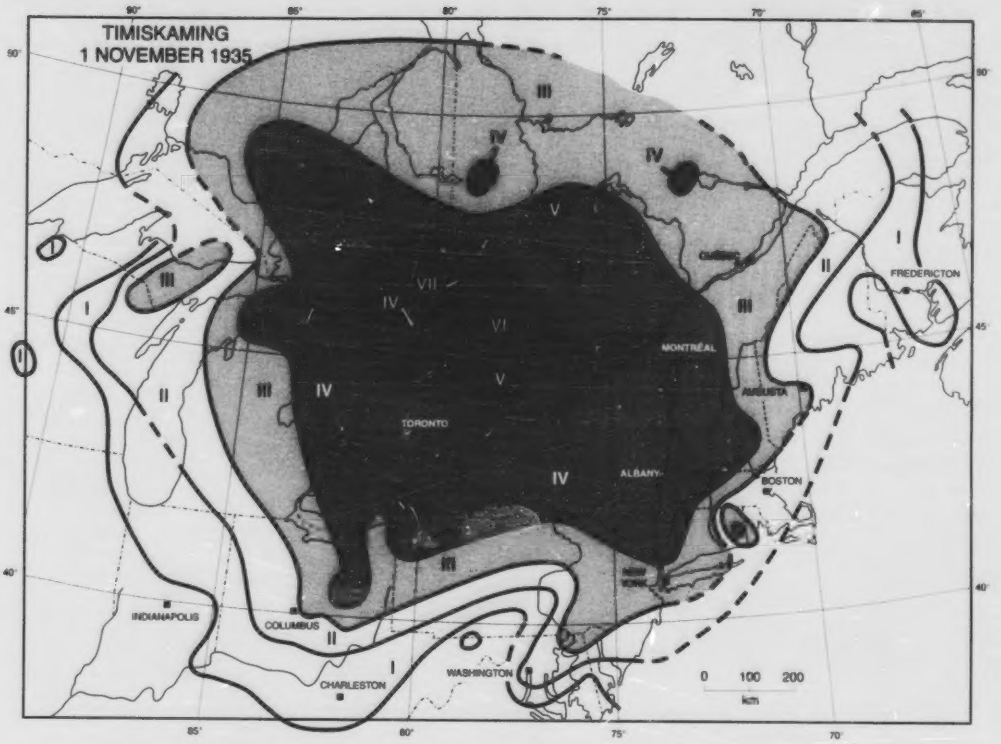

TIMISKAMING
1 NOVEMBER 1935
90°
85°
80°
75°
70°
65°
50°
45°
40°
III
IV
IV
V
VII
IV
VI
V
IV
III
III
III
II
I
II
I
IV
III
II
I
I
QUEBEC
FREDERICTON
MONTRÉAL
AUGUSTA
TORONTO
ALBANY
BOSTON
NEW YORK
INDIANAPOLIS
COLUMBUS
CHARLESTON
WASHINGTON
0 100 200
km

Figure A-5. *Isoseismal map for the 1935 Timiskaming earthquake (Modified Mercalli scale;* from *Smith, 1966).*

Figure A-6.

The 1935 Timiskaming earthquake damaged chimneys in Témiscaming, Quebec (shown here) as well as in Mattawa and North Bay, Ontario. Photograph by E.A. Hodgson. GSC 2000-075

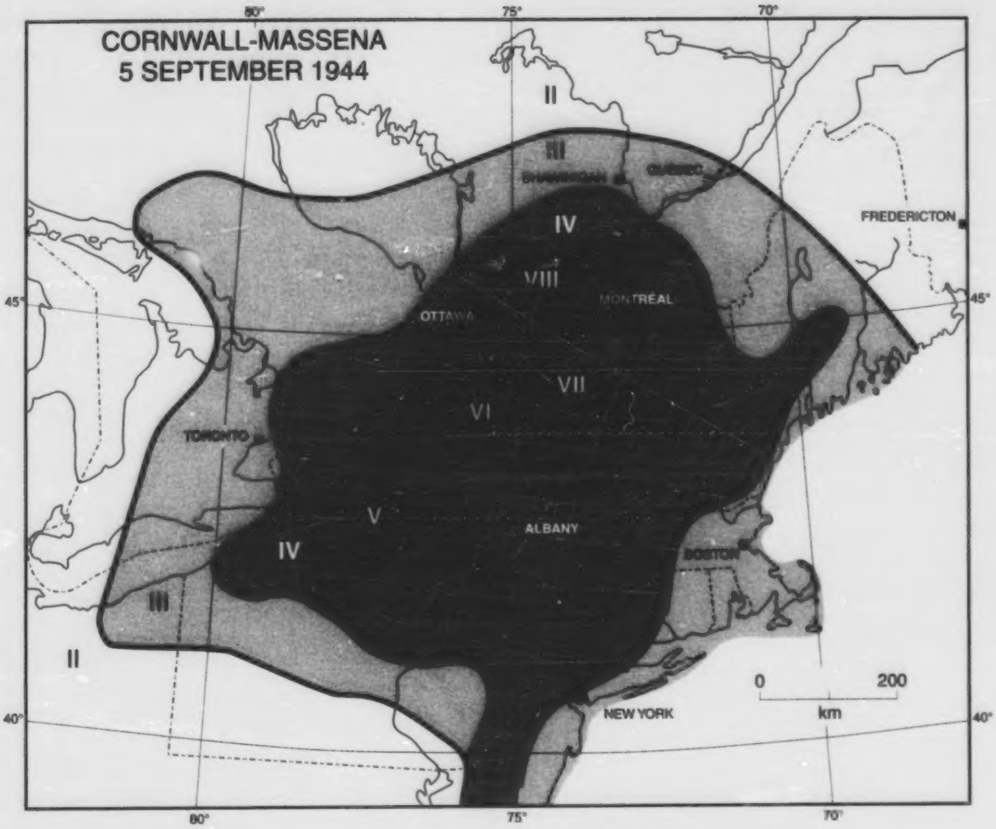

CORNWALL-MASSENA
5 SEPTEMBER 1944
80°
75°
70°
45°
40°
II
III
SHAWINIGAN
QUEBEC
IV
VIII
MONTREAL
FREDERICTON
OTTAWA
VII
VI
TORONTO
V
IV
ALBANY
BOSTON
III
II
0
200
km
NEW YORK

Figure A-7. Isoseismal map for the 1944 Cornwall earthquake (Modified Mercalli scale; from *Smith, 1966).*

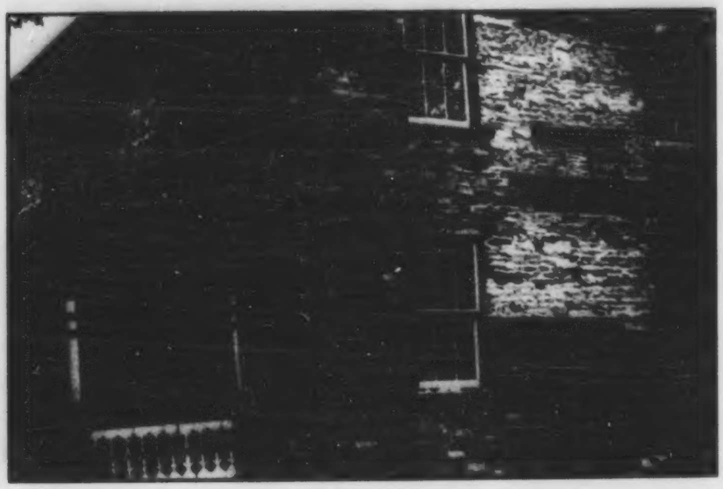


Figure A-8.

The 1944 Cornwall–Massena earthquake caused damage to unreinforced masonry buildings such as this two-story residential dwelling in Cornwall, Ontario. Photograph courtesy of C. Berkley, United States Engineer Office, New York.

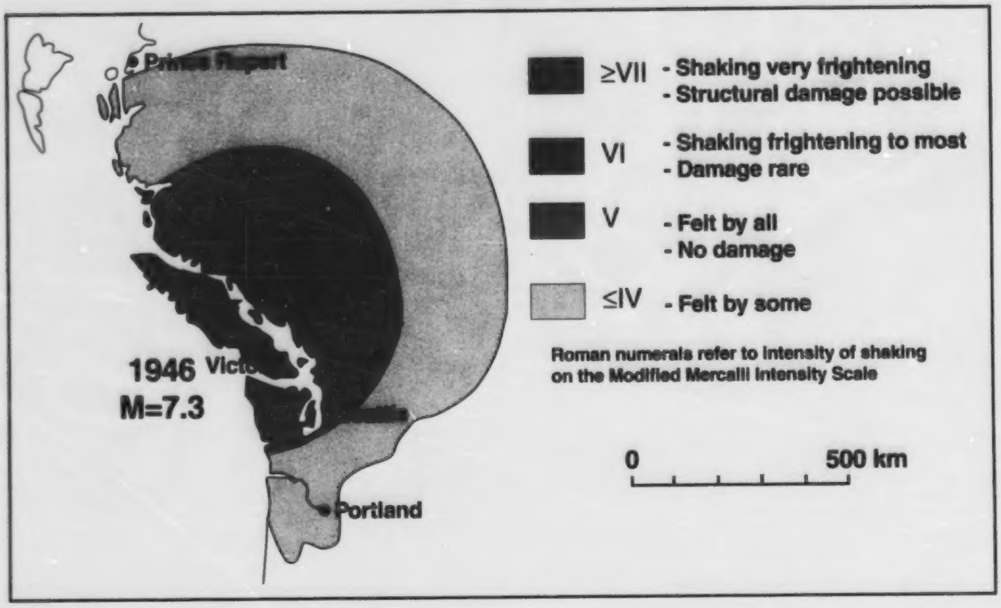

Prince Rupert
1946
M=7.3
Vic
Portland
≥VII
- Shaking very frightening
- Structural damage possible
VI
- Shaking frightening to most
- Damage rare
V
- Felt by all
- No damage
≤IV
- Felt by some
Roman numerals refer to intensity of shaking on the Modified Mercalli Intensity Scale
0
500 km

Figure A-9. *Isoseismal map for the 1946 central Vancouver Island earthquake.*

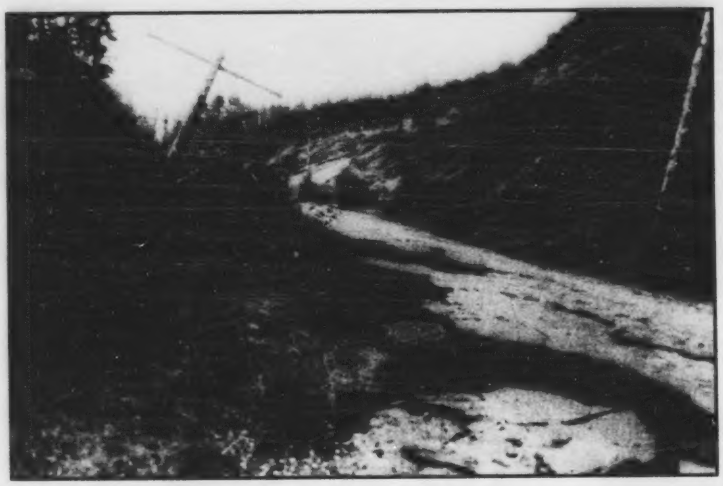


Figure A-10.

The 1946 Vancouver Island earthquake caused damage to Hern Road, near Sayward on Vancouver Island, about 60 km from the epicentre. Photograph from *the archives of the Earth Physics Branch.*

Figure A-11.

The 1946 Vancouver Island earthquake also caused building damage in Port Alberni on Vancouver Island, about 80 km from the epicentre. Photograph from *the archives of the Earth Physics Branch*

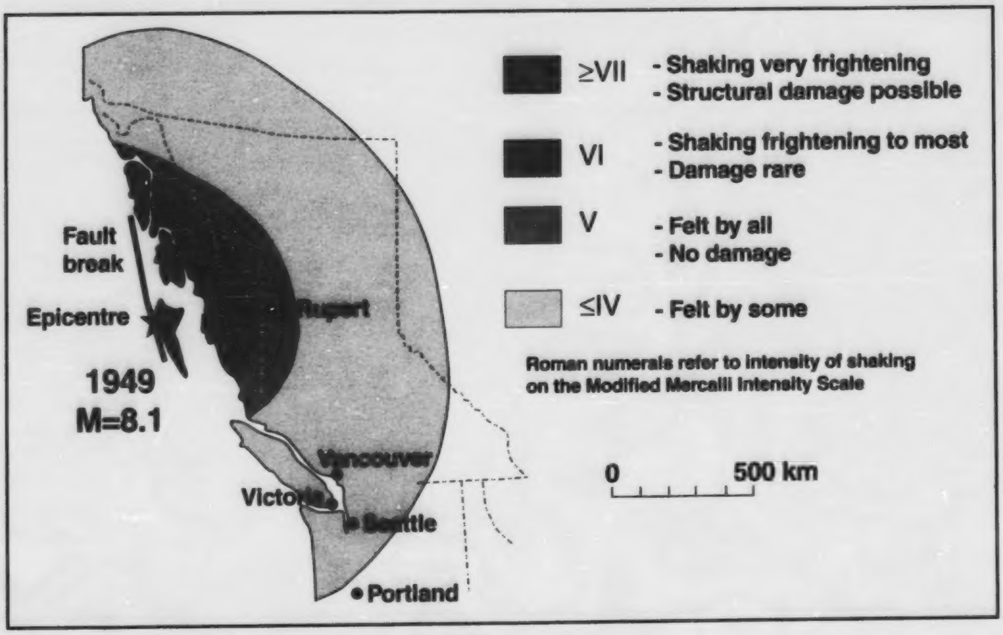

≥VII
- Shaking very frightening
- Structural damage possible
VI
- Shaking frightening to most
- Damage rare
V
- Felt by all
- No damage
≤IV
- Felt by some
Roman numerals refer to intensity of shaking on the Modified Mercalli Intensity Scale
0
500 km
Fault break
Epicentre
1949
M=8.1
Rupert
Vancouver
Victoria
Seattle
Portland

Figure A-12. Isoseismal map for the 1949 Queen Charlotte Islands earthquake.

Figure A-13.

The 1949 Queen Charlotte earthquake collapsed an oil tank at Aero, on the Queen Charlotte Islands, about 50 km east of the Queen Charlotte Fault. Photograph from *the archives of the Earth Physics Branch.*

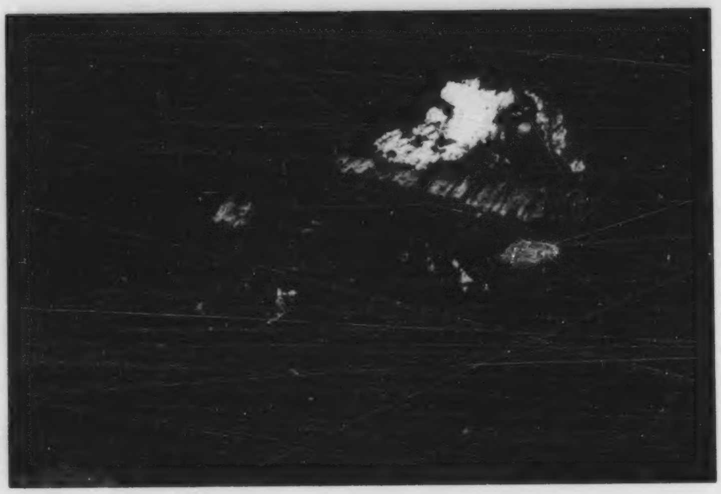


Figure A-14.

The first of two large earthquakes in the uninhabited North Nahanni river region in 1985 caused a massive rock avalanche. Photograph by S.G. Evans. GSC 2000-076

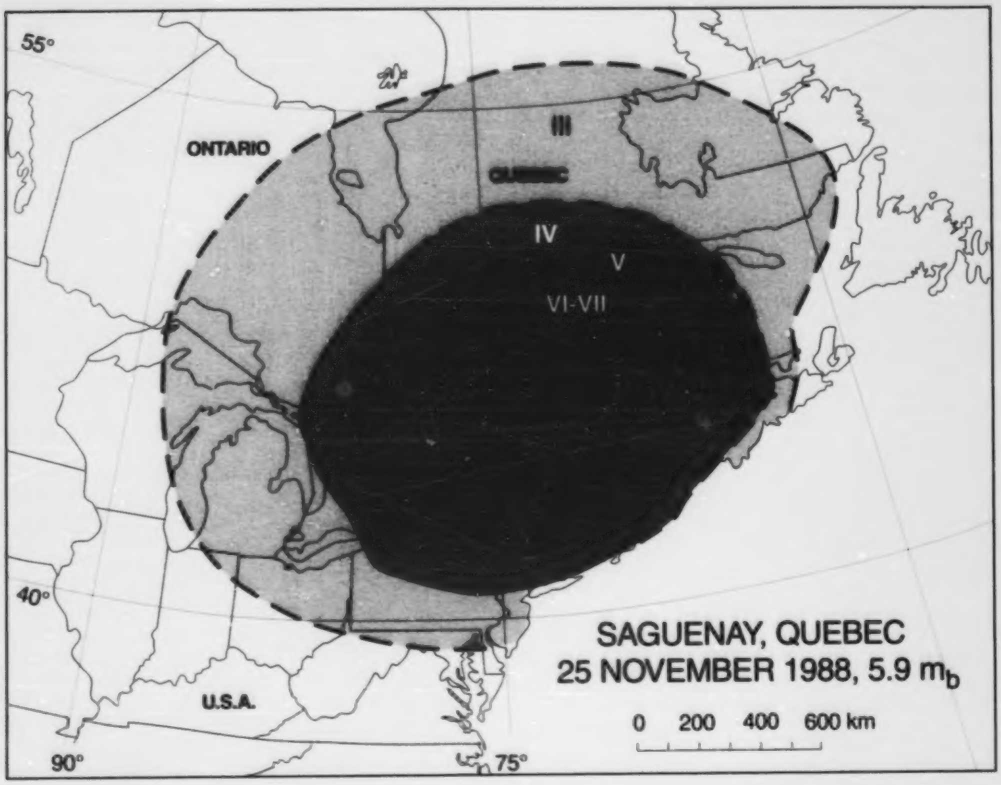

55°
ONTARIO
QUEBEC
III
IV
V
VI-VII
40°
U.S.A.
90°
75°
SAGUENAY, QUEBEC
25 NOVEMBER 1988, 5.9 m_b
0 200 400 600 km

Figure A-15. Isoseismal map for the 1988 Saguenay earthquake (Modified Mercalli scale). The total felt area of over 3.5 x 10^6 km^2 is consistent with the large amount of high-frequency energy, and approaches that of the Charlevoix–Kamouraska earthquake of 1925.

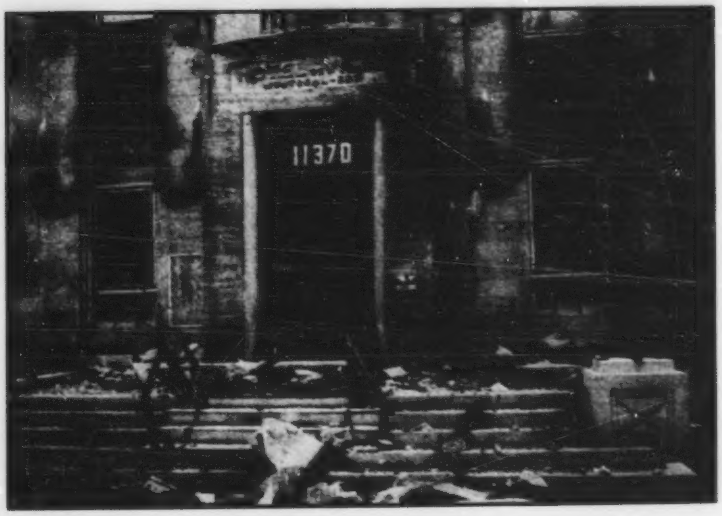


Figure A-16.

Damage from the 1988 Saguenay earthquake was reported as far away as Montréal, at the Montréal-Est City Hall (shown here), some 350 km from the epicentre. Photograph by D. Mitchell, McGill University, used by permission.

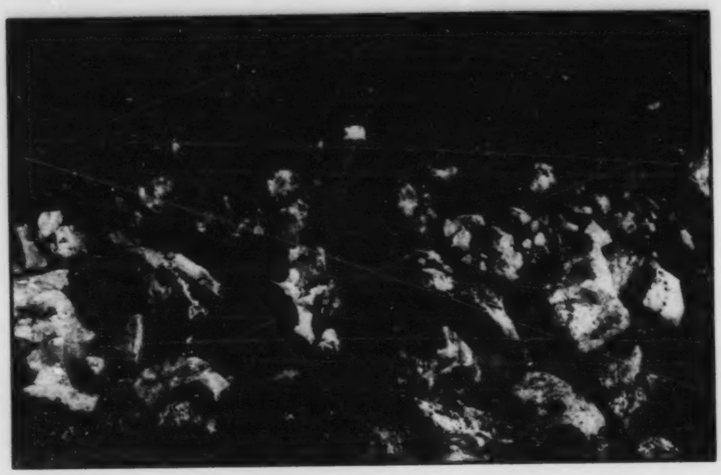


Figure A-17.

Fault scarp of the 1989 Ungava earthquake displacing a boulder field. The far side of fault was uplifted by about 2 m, most of the uplift being in this narrow band of disturbed boulders, which are light-coloured because they lack a lichen crust. GSC 2000-185

Tsunamis

John J. Clague[1]

Clague, J.J., 2001: Tsunamis; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 27–42.*

Abstract: Tsunamis are waves produced by earthquakes, landslides, volcanic eruptions, and meteorite impacts. Large tsunamis strike the British Columbia coast on average once every few hundred years. Some, including one from Alaska in 1964, are produced by distant earthquakes beneath the Pacific Ocean. Most, however, are triggered by earthquakes at the Cascadia subduction zone, which extends along the Pacific coast from Vancouver Island to northern California. Smaller, more localized tsunamis in British Columbia are triggered by landslides. Tsunamis are uncommon on the Atlantic and Arctic coasts. In 1929, however, an earthquake beneath the Grand Banks, 250 km south of Newfoundland, triggered a large submarine slump that set off a tsunami that damaged 40 communities in Newfoundland and claimed 28 lives. Tsunamis cannot be prevented, but the damage they cause can be reduced through a variety of nonstructural and structural measures, including zoning, property relocation, emergency preparedness, public education, dyking, barrier construction, floodproofing, and tsunami-resistant construction.

Résumé : Les tsunamis sont des ondes engendrées par les séismes, les glissements de terrain, les éruptions volcaniques et l'impact des météorites. Des tsunamis géants se déclenchent le long de la côte de la Colombie-Britannique en moyenne à des intervalles de quelques centaines d'années. Certains tsunamis, dont celui qui s'est produit en Alaska en 1964, sont provoqués par des séismes qui ont lieu dans des zones éloignées sous l'océan Pacifique. Cependant, la plupart des tsunamis sont provoqués par des séismes dans la zone de subduction de Cascadia, qui s'étend le long de la côte du Pacifique de l'île de Vancouver jusqu'au nord de la Californie. En Colombie-Britannique, les tsunamis localisés, de moindre importance, sont déclenchés par des glissements de terrain. Ces phénomènes se produisent rarement sur les côtes de l'Atlantique et de l'Arctique quoiqu'en 1929, un séisme survenu sous les Grands Bancs, 250 km au sud de Terre-Neuve, a provoqué un vaste glissement sous-marin qui a déclenché un tsunami, lequel a affecté 40 collectivités à Terre-Neuve et enlevé la vie à 28 personnes. Si l'on ne peut pas empêcher la formation de tsunamis, on peut en réduire les dégâts en prenant diverses mesures structurelles et non structurelles dont le zonage, la relocalisation des biens, les mesures d'urgence, la sensibilisation du public, la protection contre les inondations et la construction de digues, de barrages et de structures pouvant résister aux tsunamis.

---

[1] GSC Vancouver

WHAT IS A TSUNAMI?

Tsunamis are ocean waves generated by underwater disturbances of the seafloor (Fig. 1) or by surface impacts. They are triggered by earthquakes and, less commonly, by landslides, volcanic eruptions, and meteorite impacts. A large meteorite that strikes the sea could produce waves hundreds of metres high, but the probability of such an event is extremely low. Earthquake-triggered 'tsunamis', which in Japanese mean 'harbour waves', are also called 'seismic sea waves' and, erroneously, 'tidal waves'. They are imperceptible on the open ocean, where they have amplitudes of less than 1 m and move at up to 1000 km/h, but can reach heights of up to 30 m or more in shallow water and cause major damage to coastal areas (Fig. 2). When a tsunami approaches the shore, shoaling causes waves to decelerate and grow taller, with the distance between them decreasing.

A popular misconception about tsunamis is that a single, very large wave breaks onto the shore. A tsunami, however, is not a single breaker, but rather a series of waves separated by minutes to an hour or more. Generally, the second or third wave is the largest. The waves commonly are turbulent, onrushing surges and not simple breakers. When one wave overtakes another, however, a steep wall of water, or bore, may be created. In some areas, a tsunami may be less energetic, with water levels rising rapidly, but no strong landward surge of water.

Wave runup is dependent on offshore bathymetry, shoreline orientation and shape, onshore topography, and perhaps other factors. Because these parameters vary along most coasts, runup may differ considerably over even short distances, making it difficult to assess local hazard. In general, runup is small on straight, steep shorelines and high at the heads of shallow, broad bays and some inlets.

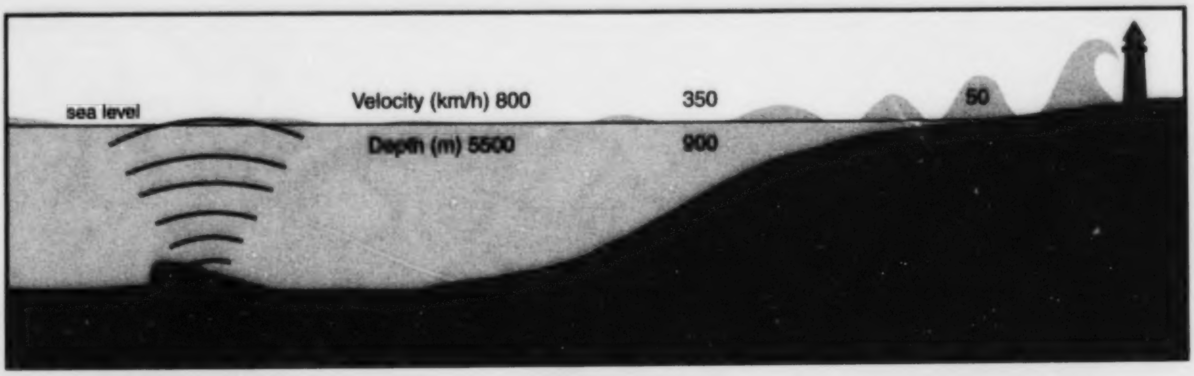


Figure 1. A large earthquake displaces the ocean floor, producing a tsunami that propagates outward from the source. The velocity of the tsunami is proportional to the square root of the ocean depth at any given point — as shown in this Figure, the velocity of a wave in 5500 m of water is over 800 km/h, but in 20 m of water, it is only 50 km/h. The waves become higher and their spacing shorter as they approach the shore.

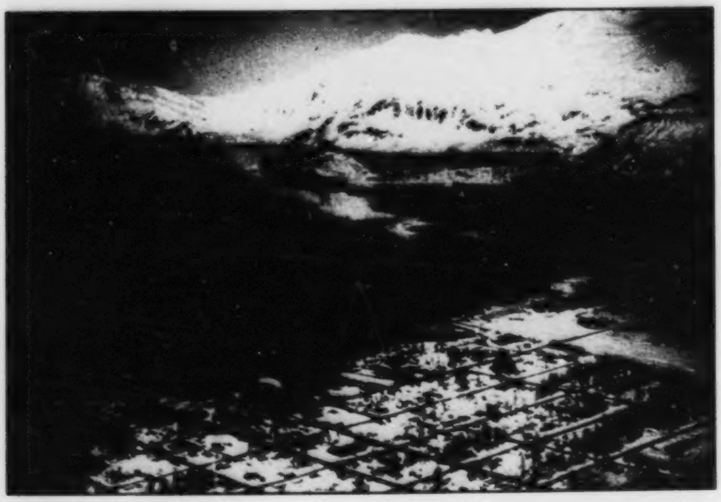


Figure 2.

Aerial view of Valdez, Alaska, showing the extent of tsunami inundation during the 1964 Alaska earthquake. A large submarine slide destroyed the dock area and part of the town and generated a wave that slammed into the waterfront within two or three minutes of the beginning of the earthquake. This wave demolished what was left of the waterfront, destroyed the fishing fleet, and penetrated about two blocks into the town. Property damage at Valdez was $15 million (1964 U.S. dollars) and there were 30 fatalities. (United States Geological Survey photograph, courtesy of the National Oceanic and Atmospheric Administration)

TSUNAMI DISASTERS

The main sources of destructive tsunamis in the twentieth century have been earthquakes at the Nankai Trough and Kurile Trench east of Japan, the Aleutian Trench south of Alaska, and the Peru-Chile Trench west of South America (Fig. 3). Two examples serve to illustrate the devastation that can be wrought by earthquake-triggered tsunamis. In 1896, waves up to 35 m high struck the east coast of Japan, smashing more than 100 000 houses and drowning 27 000 people. In 1946, an earthquake with a Richter magnitude (M) of 7.3 near Unimak Island, Alaska, generated a tsunami that was strong enough to wipe out both the five-story Scotch Cap Lighthouse, situated almost 10 m above sea level (Fig. 4), and a nearby radio antenna, 35 m above the sea. Although this tsunami had surprisingly little effect elsewhere in Alaska, it

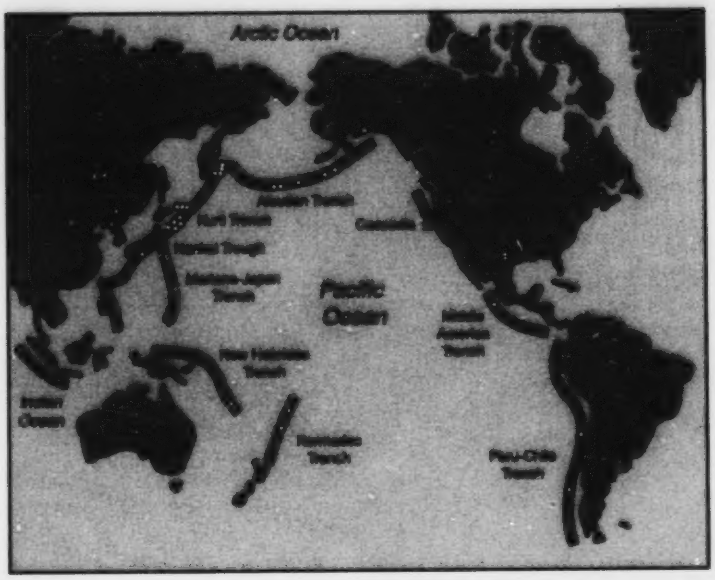

Arctic Ocean
Aleutian Trench
Kuril Trench
Cascadia
Nankai Trough
Marianas-Japan Trench
Pacific Ocean
Middle America Trench
New Hebrides Trench
Indian Ocean
Kermadec Trench
Peru-Chile Trench

Figure 3.

Most large tsunamis in the Pacific Ocean have been triggered by great earthquakes at the subduction zones (black bands) shown here. White dots locate the sources of tsunamis larger than 10 cm in height that were recorded at the Tofino tide gauge on the west coast of Vancouver Island between 1905 and 1981 (Table 1).

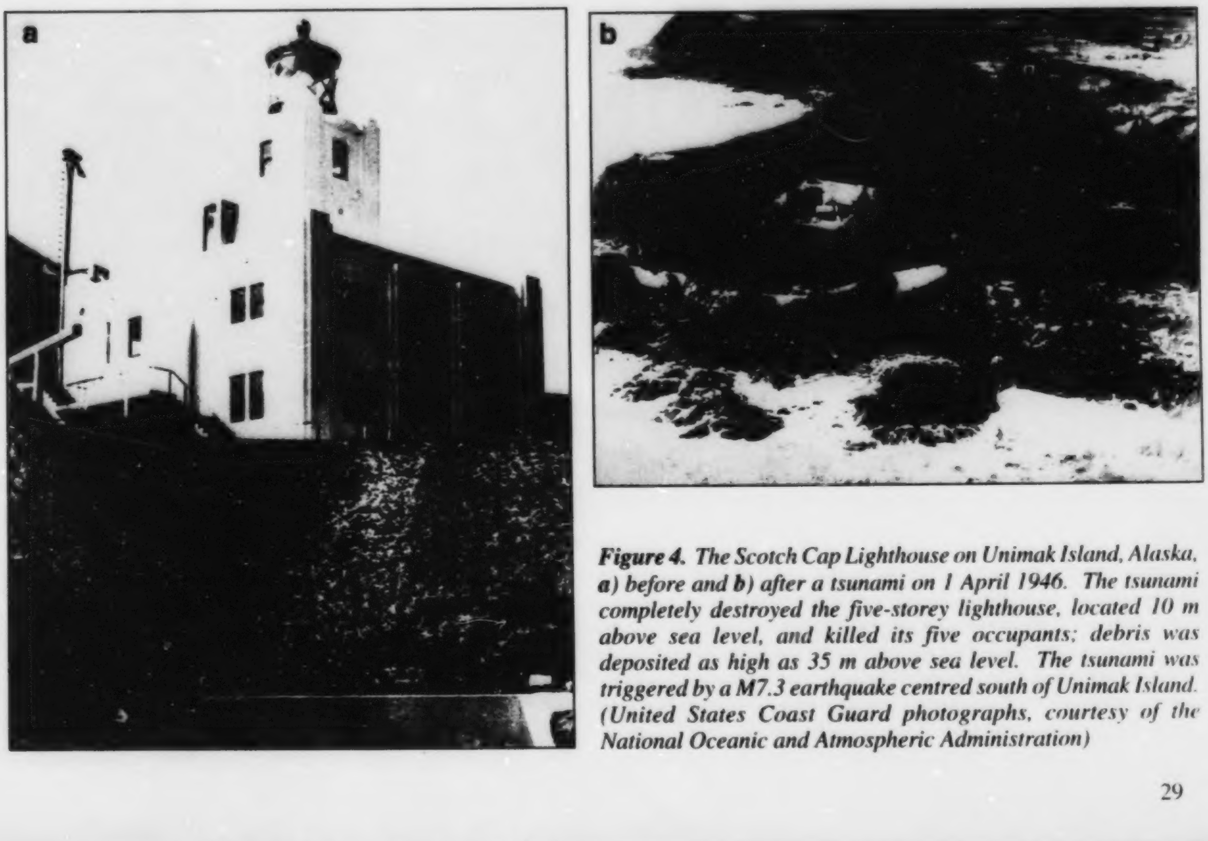


Figure 4. *The Scotch Cap Lighthouse on Unimak Island, Alaska, **a**) before and **b**) after a tsunami on 1 April 1946. The tsunami completely destroyed the five-storey lighthouse, located 10 m above sea level, and killed its five occupants; debris was deposited as high as 35 m above sea level. The tsunami was triggered by a M7.3 earthquake centred south of Unimak Island. (United States Coast Guard photographs, courtesy of the National Oceanic and Atmospheric Administration)*

was one of the most destructive to ever strike the Hawaiian Islands, killing 159 people and causing millions of dollars in damage. However, the most destructive historical tsunami was caused by a volcanic eruption, not an earthquake. The explosive eruption of the island of Krakatoa in 1883 triggered huge waves that killed 36 500 people in Indonesia.

HISTORICAL TSUNAMIS IN CANADA

Pacific coast

The largest historical tsunami on the west coast of Canada was triggered by the great (M9.2) Alaska earthquake of 27 March 1964. A series of waves moved radially outward from the epicentre of the earthquake near the head of Prince William Sound and, within a few hours, reached the outer coasts of the Queen Charlotte Islands and Vancouver Island (Fig. 5), causing about $10 million in damage (1964 dollars), mainly to the communities of Port Alberni, Hot Springs Cove, and Zeballos.

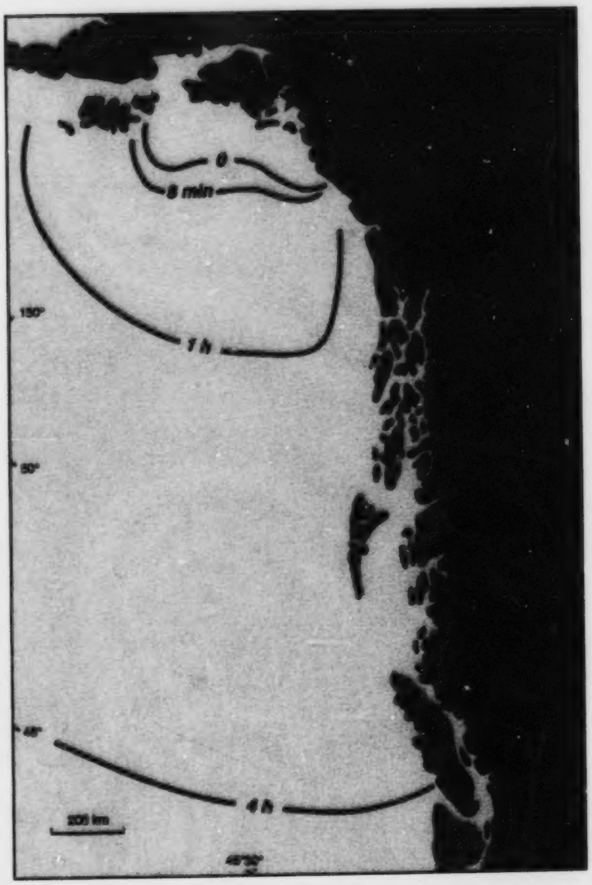


Figure 5. Travel times of the 1964 tsunami from the earthquake epicentre in south-central Alaska.

Fortunately, tsunamis comparable to those generated by the 1964 earthquake are rare. Of the 176 tsunamis recorded in the Pacific Ocean between 1900 and 1970, 35 caused damage near their sources, but only 9 resulted in widespread destruction, and none other than the 1964 event significantly damaged British Columbia. Only 2 of the 43 tsunamis registered at Tofino, on the west coast of Vancouver Island, between 1906 and 1981 produced waves more than 1 m high — the 1964 tsunami, with a maximum wave height (trough to crest) of 2.40 m, and a tsunami in 1960 generated by the great Chilean earthquake, with a wave height of 1.26 m (Table 1).

Some tsunamis in British Columbia are products of local earthquakes and submarine landslides. These locally generated waves can damage areas that would be unaffected by tsunamis of distant earthquakes. For example, sea waves triggered by a large submarine landslide at the head of Douglas Channel in 1975 slammed into shore installations at Kitimat, causing about $600 000 in damage. A small tsunami was produced by the Vancouver Island earthquake of June 1946; it affected shores along the Strait of Georgia and nearby inlets. The main wave had a maximum amplitude of over 2 m; a slightly smaller secondary wave was responsible for the death of one person (the only casualty of the earthquake) near Mapleguard Point on eastern Vancouver Island.

Table 1. Tsunamis recorded at Tofino, British Columbia.

Date	Source area [1]	Height of maximum wave (cm) [2]
1 November 1915	Japan	12
1 May 1917	Kermadec Islands	12
18 July 1918	Indonesia	16
7 September 1918	Kuril Islands	16
30 April 1919	Tonga	15
11 November 1922	Peru	27
3 February 1923	Kamchatka	27
13 April 1923	Kamchatka	15
7 March 1929	Aleutian Islands	11
2 March 1933	Japan	23
30 November 1934	Mexico	22
10 November 1938	Aleutian Islands	27
7 December 1944	Japan	12
27 December 1944	New Hebrides (Vanuatu)	12
1 April 1946	Aleutian Islands	58
4 March 1952	Japan	12
4 November 1952	Kamchatka	58
9 March 1957	Aleutian Islands	52
11 March 1957	Aleutian Islands	18
6 November 1958	Kuril Islands	10
22 May 1960	Chile	126
13 October 1963	Kuril Islands	16
28 March 1968	south-central Alaska	240
16 May 1968	Japan	12
16 May 1968	Japan	13

Source: Wigen (1983, Appendix B); only tsunamis with maximum wave heights of 10 cm or more are included.

[1] Foci of earthquakes are beneath the Pacific Ocean near the named areas (Fig. 2).

[2] Sum of displacements of maximum wave and successive trough from corresponding tide levels.

The 1964 tsunami

The Alaska earthquake of 27 March 1964, the second largest (M9.2) of the twentieth century, triggered tsunamis that killed 130 people, some as far away as California. The main tsunami swept southward across the Pacific Ocean at about 830 km/h, reaching Antarctica in only 16 hours. Waves were amplified as they moved up Barkley Sound and Alberni Inlet on Vancouver Island; the waves were 2.5 times larger at Port Alberni, at the head of the inlet, than at Tofino and Ucluelet on the open coast.

Three main waves struck Port Alberni between 00:20 h and 03:30 h on 28 March (Fig. 6). Most people in the town were asleep when the first wave rolled in. The sea surged up Somass River at about 50 km/h and spilled onto the land, inundating whole neighbourhoods with chest-deep water. This first wave reached 3.7 m above geodetic datum (geodetic datum corresponds approximately to mean sea level or zero elevation) and knocked out the Port Alberni tide gauge.

The second and most destructive wave swept into town less than two hours later, at 02:00 h. The lights of the waterfront mills went out as the water smashed through the facilities. The ground floor of the Barclay Hotel, 1 km inland, was splintered by the surging water; guests had to be rescued from an upper floor by police in boats. Logs and debris crashed into buildings, automobiles were cast about, and houses were swept off their foundations and hurtled inland. As the water subsided, some buildings were dragged seaward; two houses drifted into Alberni Inlet and were never seen again. The second wave left a mark on the tide station at 4.3 m above geodetic datum.

The third wave, which arrived at about 03:30 h, was the largest of all, but because the tide had fallen, it crested at 3.9 m and did little additional damage. Other waves oscillated in Alberni Inlet, with decreasing strength, for two more days.

Two hundred and sixty houses in Port Alberni were damaged by this tsunami, 60 extensively, and the total economic loss in the town was about $5 million (1964 dollars).

The 1964 tsunami caused damage elsewhere on Vancouver Island — it sank a fishing boat and damaged log booms at Ucluelet, damaged wharf facilities at Tofino, breached the municipal water pipeline that crosses the seafloor near Tofino, destroyed an aboriginal village at Hot Springs Cove, and swept buildings off their foundations at Zeballos. Waves were largest at the heads of some inlets due to focusing of wave energy and resonance effects.

Atlantic coast

Tsunamis in the Atlantic Ocean are generated by crustal earthquakes and submarine landslides. Most Atlantic tsunamis are destructive only near their sources; in theory, however, a destructive tsunami might be generated by a distant, huge, submarine landslide. There has been only one destructive historical tsunami in Atlantic Canada, but this event claimed many more human lives than any other historical tsunami, or for that matter historical earthquake, in Canada.

The 1929 tsunami

On Monday, 18 November 1929, a M7.2 earthquake occurred about 20 km beneath the seafloor at the southern edge of the Grand Banks, 250 km south of Newfoundland. The earthquake was felt as far away as New York and Montréal, but caused no damage on land because of its remoteness; in Newfoundland, windowpanes rattled and dishes fell from cabinets. The shaking, however, triggered a huge submarine slump on the continental shelf. The slump transformed into a turbid flow of clay, silt, and sand that moved at up to 100 km/h into deeper water, breaking 12 telegraph cables in its path. Some 200 km^3 of material were redistributed over an area of 150 000 km^2 by the slump and turbidity current.

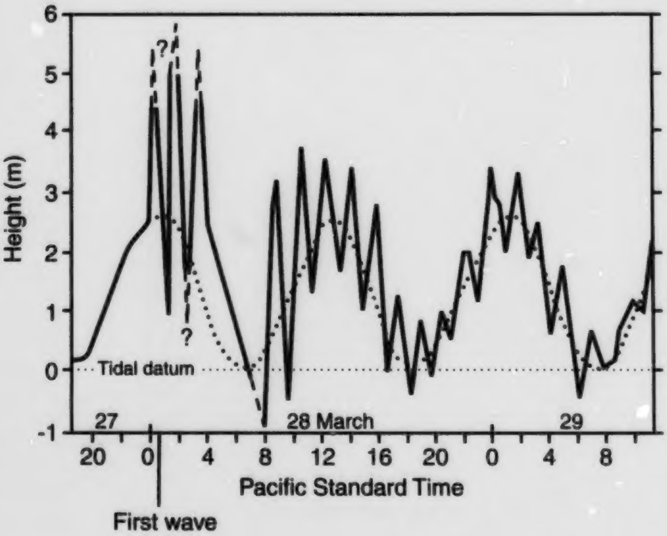


Figure 6.

Tidal records for Port Alberni, British Columbia, 27–29 March 1964. The three main waves of the Alaska tsunami arrived between 00:20 h and 03:30 h on 28 March, forcing the tide gauge offscale; normal tidal heights are shown by a dotted line (modified from *Wigen and White, 1964*).

The slump set off a tsunami that propagated across the Atlantic (Fig. 7), registering on tide gauges as far away as South Carolina and Portugal; however, it did no damage at these distant locations. The tsunami arrived in Bermuda about two hours after the earthquake, travelling on average at just over 700 km/h. In contrast, it took about 2.5 hours to travel the much shorter distance to the Burin Peninsula of Newfoundand (Fig. 7), because of the drag exerted by the shallow continental shelf north and west of the earthquake epicentre.

The Burin Peninsula on Newfoundland's south coast was hardest hit by the tsunami (Fig. 8). Three main waves thundered up narrow channels and into bays over a half-hour period on the evening of November 18. The first wave was preceded by an unusual withdrawal of water that left anchored boats in coastal communities high and dry. In some areas, this withdrawal was followed by a turbulent onrush of water, probably a bore. One eyewitness, then a young boy, described the wave coming in smooth and fast; with the bright moon shining on its crest, it looked like a car with its headlights beaming, driving up the harbour. In other areas, where there was relatively deep water near the shore, the sea rose very rapidly with little turbulence. The tsunami lifted small boats and schooners 5 m high, snapping anchor chains and tossing the craft onshore or engulfing them. Houses floated from their foundations; some were splintered, whereas others were swept back and forth by the flooding and ebbing waters (Fig. 9). The tsunami transported sand and gravel as much as 1 or 2 km inland; traces of the deposit are still preserved in many areas (Fig. 10). More than 40 coastal communities on Burin Peninsula were damaged and the tsunami claimed 27 lives in Newfoundland and 1 in Nova Scotia. Property damage, including the cost of repairing the undersea cables, was $400 000 (1929 dollars). Damage was made worse by the fact that the tsunami arrived near the peak of a very high tide (Fig. 11). Maximum wave heights on the Burin Peninsula are not known; estimates at the time of the disaster ranged from 9 to 15 m, but recent field investigations suggest lower values — about 4.6 m in Burin and Port au Bras, 3 m in Lamaline and Point au Gaul, and perhaps 7.5 m in Taylor's Bay. Surging waters may have reached higher elevations at some sites.

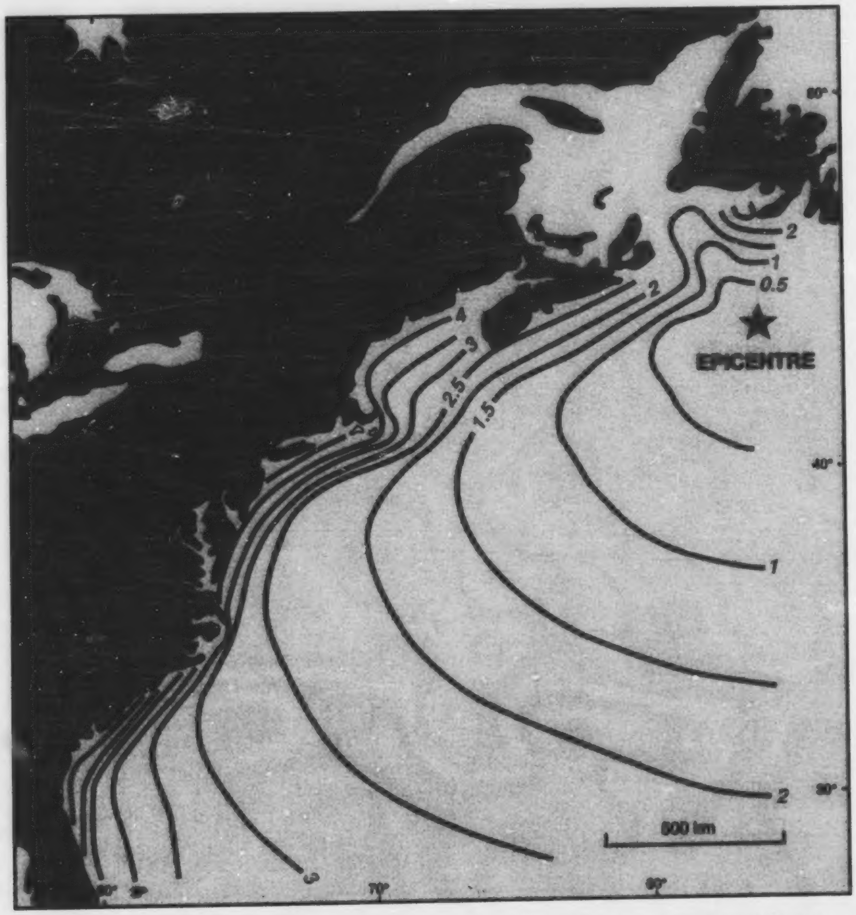


Figure 7. Travel times of the 1929 tsunami from the epicentre of the Grand Banks earthquake.

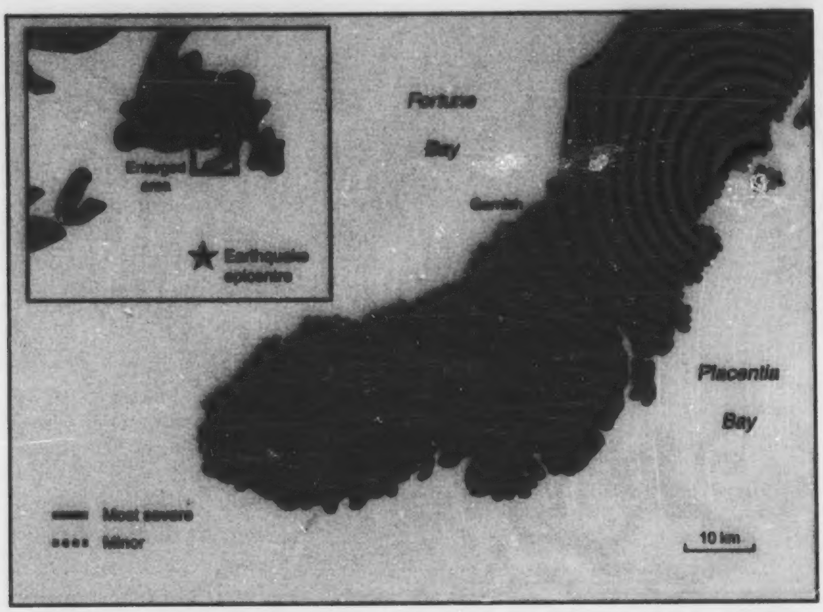

Enlarged area
Earthquake epicentre
Fortune Bay
Garnish
Placentia Bay
Most severe
Minor
10 km

***Figure 8.** Extent of damage from the 1929 tsunami on Burin Peninsula, Newfoundland* (modified from *Whelan, 1994).*

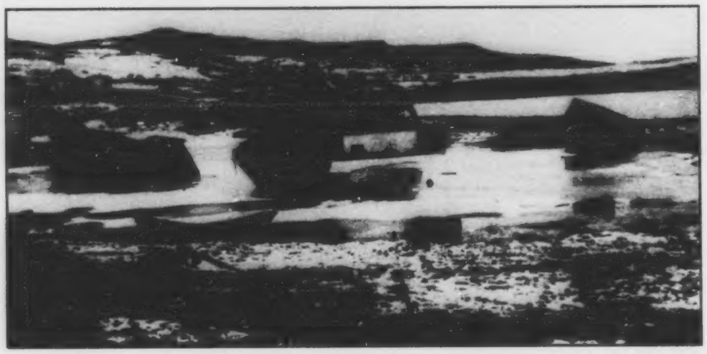


Figure 9.

Coastal communities on Burin Peninsula bore the brunt of the 1929 tsunami. This photograph shows buildings in Lord's Cove that were tossed and smashed by the tsunami (photograph by H.M. Mosdell, from the collection of W.M. Chisholm; courtesy of Alan Ruffman).

Figure 10.

Layer of tsunami-laid sand and fine gravel exposed in a shallow excavation at Taylor's Bay, Newfoundland. The layer is overlain and underlain by peat. It averages about 6 cm in thickness in this exposure, but reaches up to 25 cm thick elsewhere in the area. Note how the sand thins over a boulder near the left side of the photo (photograph by M. Tuttle, courtesy of T. Anderson).

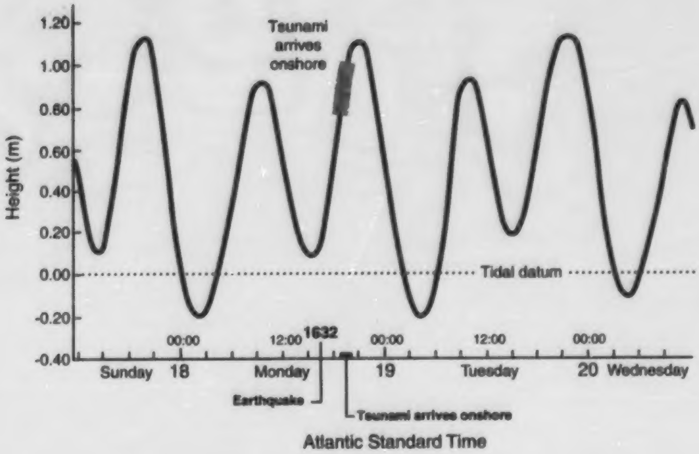


Figure 11.

The devastation and loss of life from the 1929 tsunami were made worse by the fact that the waves struck Burin Peninsula near the peak of a very high ('spring') tide. The tidal curve shown here is for North Sydney, Nova Scotia, and is based on hindcast astronomical prediction (Canadian Hydrographic Service, Bedford Institute of Oceanography).

PREHISTORIC TSUNAMIS IN CANADA

Geological evidence

Photographs and eyewitness accounts of historical tsunamis in Chile, Hawaii, Indonesia, Japan, and elsewhere indicate that large amounts of sediment are transported and deposited by the turbulent, landward-surging waves. Most of the sediment is derived from loose nearshore deposits, generally sand and silt, but in some cases gravel. The sediment is carried by the turbulent water in suspension and by traction. It is deposited as sheets in the upper part of the intertidal zone and in low-lying areas inland of the shore as the waves lose energy or begin to recede.

Tsunami sand sheets have been found in mud and peat deposits in protected tidal marshes at several localities on western Vancouver Island (Fig. 12) and in Washington and Oregon. They are evidence that large tsunamis have struck the Pacific coast on many occasions during the last few thousand years. The sand sheets have considerable areal extent and sharp contacts with overlying and underlying deposits. In most cases, they consist of massive sand and silty sand with abundant wood and other plant detritus. Locally, however, the deposits are gravelly, graded, or consist of two or more paired laminae (couplets) of silt and sand. The sands commonly contain marine microfossils and thin, fine, and rise landward, indicating that they were deposited by landward surges of water and not by streams.

In open coastal settings, the deposits of rare large storms are commonly difficult to distinguish from tsunami deposits. Tsunamis, however, typically spread sediment widely above the upper limit of tides, whereas storm waves are less likely to move sediment far inland. Storm deposits are unlikely to be found far from the shore in protected tidal inlets because, as water levels rise, current velocities in channels drop below values required to transport enough sand to form widespread sheets. Also, some sand sheets consist of two or more fining-upward beds, suggestive of deposition by successive tsunami waves rather than storm surges.

Figure 12. *Tsunami sand (S) sharply overlying a peaty soil (SO) in a pit at a tidal marsh near Tofino, British Columbia. The sand is abruptly overlain by tidal mud (M), which grades upward into peat (P) of the present-day marsh. The tsunami that deposited the sand was triggered by a great earthquake at the Cascadia subduction zone 300 years ago. Photograph by J.J. Clague. GSC 2000-077*

Other evidence supports the idea that these anomalous tidal-marsh sand sheets are tsunami deposits. Some sand sheets directly overlie buried peaty or forest soils that subsided during large earthquakes (Fig. 12). The subsidence was caused by the sudden release of strain in the Earth's crust during earthquakes. In places, the sands entombed delicate leaves and stems of herbaceous plants that were growing on the marsh before it subsided; this shows that sand deposition occurred immediately after subsidence and implies that the two events are linked, the only plausible common cause being a large earthquake.

Buried soils resulting from episodic, sudden subsidence are present at over a dozen estuaries on the Pacific coast between central Vancouver Island and northern California. Several lines of evidence show that the buried soils subsided during earthquakes: 1) the soils were submerged suddenly, rather than gradually; 2) the amount of submergence in many areas is more than 0.5 m and ranges up to 2 m; 3) the soils are laterally extensive; 4) submergence events at widely spaced sites are the same age; and, as noted above, 5) submergence coincided with tsunami deposition. This evidence further indicates that the earthquakes were subduction, or plate-boundary, events that involved rupture of tens to hundreds of kilometres of the boundary between the subducting oceanic Juan de Fuca Plate and the overriding continental North American Plate (a region referred to as the 'Cascadia subduction zone'). Analogues for these prehistoric earthquakes include the great subduction earthquakes in Chile in 1960 and in Alaska in 1964, both of which produced widespread crustal subsidence in the source areas and destructive tsunamis that crossed the Pacific Ocean.

Another argument for a tsunami origin for the sand sheets is that the uppermost sheet at many marshes on Vancouver Island has been shown to have been deposited by the 1964 Alaska tsunami. Older sand sheets at these sites are similar in geometry and physical properties to the 1964 deposit and thus, by analogy, were probably also emplaced by tsunamis. Some of the older sand sheets, however, are thicker and coarser grained than the 1964 deposit, suggesting that the older tsunamis were larger than the 1964 event.

Recently, tsunami sands have been found in coastal lakes in Oregon and on western Vancouver Island. Tsunamis enter these lakes by surging up outlet streams or by crossing sand dunes that lie between some of the lakes and the sea. The sands are much coarser than the organic-rich, silty and clayey sediments that accumulate in the lakes under normal conditions and they commonly contain abundant plant detritus. The most complete record found to date is from Bradley Lake in Oregon, where there are 14 tsunami layers in a sequence spanning the last 7500 years.

The evidence for prehistoric tsunamis has come largely from the outer coast and from inlets and fiords on Vancouver Island; however, a tsunami sand sheet has also been found along the shores of Puget Sound and nearby Hood Canal. An earthquake on a fault near Seattle, Washington, about 1000 years ago, displaced the floor of Puget Sound and triggered a tsunami that propagated north and south from the earthquake epicentre. This tsunami deposited a sheet of sand on tidal peats at many marshes in the area. Researchers have studied the thickness, texture, and spatial distribution of the sand at Cultus Bay, Whidbey Island, 35 km from the epicentre, and concluded that the tsunami moved across the marsh at this site at a velocity of 6 m/s, inundating it to a depth of about 1 m. The sand was eroded from tidal flats seaward of the marsh and transported inland by the tsunami. Notably, no evidence exists for other tsunamis during the past 2000 years either at this marsh or at any other where the 1000 year old sand sheet has been found. This indicates that large tsunamis are rare events in protected inland waterways such as Puget Sound. No tsunami deposits have yet been identified along the shores of the Strait of Georgia.

So far, evidence for prehistoric tsunamis has been found only on the west coast, but it is likely that similar evidence exists in Atlantic Canada. A widespread layer of 8000 year old marine sediment in coastal areas of eastern Scotland, western Norway, and Iceland was probably deposited by a large tsunami that was triggered by a submarine landslide in the North Atlantic Ocean. On the basis of the known extent and character of the deposit, the tsunami may have been large enough to have left a trace on the far side of the Atlantic, in Canada. It is also possible that other submarine landslides or large earthquakes, such as the Lisbon, Portugal, earthquake of AD 1755, have triggered large tsunamis that have struck Atlantic Canada.

Native oral traditions

Oral traditions of Northwest Coast aboriginal peoples tell of large earthquakes and tsunamis. An example is the story told to Judge James Swan, a well known student of native lore, in 1868 by a chief of aboriginals living at Cape Flattery in northwestern Washington, just south of Vancouver Island:

> ...The water suddenly receded leaving Neeah Bay perfectly dry. It was four days reaching its lowest ebb, and then rose again without any waves or breakers, till it had submerged the Cape, and in fact the whole country, excepting the tops of mountains at Clayoquot. The water on its rise became very warm, and as it came up to the houses, those who had canoes put their effects in them, and floated off with the current, which set very strongly to the north. Some drifted one way, some another; and when the waters assumed their accustomed level, a portion of the tribe found themselves beyond Nootka, where their descendants now reside...Many canoes came down in the trees and were destroyed, and numerous lives were lost. The water was four days regaining its accustomed level. (Swan, 1868; *in* Heaton and Snavely, 1985, p. 1457).

Obviously, no tsunami could have the height (about 400 m) or duration (four days) of the event reported in the Cape Flattery legend. This raises the question of whether the account might be entirely fictional. However, the description of the water receding and then returning in a strong current is suggestive of a tsunami. In addition, the account is very similar to an oral tradition from the Port Alberni area of a great ebb and flow of the sea. Thus, the event probably happened, although it is greatly exaggerated in the oral traditions. In this regard, James Swan separated the Cape Flattery account from the majority of legends that he felt were of a more mythical nature. He reported that the event occurred "a long time ago but not at a very remote period"; it thus predated the time of the aboriginal who reported it to him, but was not of an ancient, mythical era.

There is nothing in this either the Cape Flattery or Port Alberni oral traditions to indicate the source of the presumed tsunami(s). However, the following account suggests that ground shaking may have immediately preceded a tsunami that destroyed Loht'a, the principal winter village of the

Pachena Bay people of western Vancouver Island, sometime prior to European contact. If so, this tsunami must have been generated by an earthquake off Vancouver Island or Washington, as opposed to a distant, large earthquake elsewhere in the North Pacific.

> This story is about the first Anaqtl'a or 'Pachena Bay' people. It is said that they were a big band at the time of him whose name was H̲ayoqwis is, 'Ten-On-Head-On-Beach'. He was the Chief; he was of the Pachena Bay tribe; he owned the Pachena Bay country. Their village site was Loh̲t'a; they of Loh̲t'a live there. I think they numbered over a hundred persons...There is no one left alive due to what this land does at times. They had no way or time to try to save themselves. I think it was at nighttime that the land shook...They were at Loh̲t'a:, and they simply had no time to get hold of canoes, no time to get awake. They sank at once, were all drowned; not one survived...I think a big wave smashed into the beach. The Pachena Bay people were lost...But they on their part who lived at Ma:lts'a:s, 'House-Up-Against-Hill', the wave did not reach because they were on high ground. Right against a cliff were the houses on high ground at M'a:lsit, 'Coldwater Pool'. Because of that they came out alive. They did not drift out to sea along with the others.... (Arima et al., 1991, p. 230–231)

Recurrence

The history of large subduction earthquakes and attendant tsunamis in the Pacific Northwest has been inferred by dating suddenly buried soils and overlying tsunami sand sheets at many sites along the Pacific coast from northern California to Vancouver Island. Fossil plants that were killed by submergence or tsunamis, or that became established at about the same time as these events, have been dated using radiocarbon and dendrochronological (tree-ring counting) techniques. Earthquake ages have also been estimated by measuring thicknesses of clay beds separating seismically triggered turbidites in deep-sea cores collected off the coasts of Washington and Oregon.

Previous estimates of average recurrence intervals for tsunami-generating earthquakes along the Cascadia subduction zone range from 200 to 600 years, but these estimates have uncertainties that may total many hundreds of years. The most recent and best estimate of average recurrence, about 500 years, comes from a detailed study of well exposed buried soils at estuaries in southwestern Washington State. Seven large earthquakes have struck this area in the last 3500 years. The age ranges for five of the seven events were determined from high-precision (long-count-time) radiocarbon ages on plants that died, or began to grow, at about the time of the earthquake. The age ranges for the other two events are larger and were established from conventional radiocarbon ages and stratigraphic position.

Intervals between successive earthquakes in southwestern Washington State differ considerably. Three earthquakes occurred between 3320–3500 and 2400–2800 years ago; their mean recurrence is thus more than 270 years, but less than 550 years. An interval of 700 to 1300 years separated the youngest of these three events from the next recognized earthquake, which dates to sometime between 1500 and 1700 years ago. Two earthquakes occurred 1130 to 1350 and 900 to 1300 years ago; after this, a period of 600 to 1000 years elapsed before the most recent earthquake, 300 years ago.

The most recent of the seven earthquakes documented in southwestern Washington State is also the best dated and evidence for it has been found along the full 900 km length of the Cascadia subduction zone from northern California to Vancouver Island. The possible age range, in this instance, has been minimized by dating large numbers of earthquake-killed plants rooted in the uppermost buried soil and by using high-precision radiocarbon methods that provide ages with uncertainties of 10 to 20 years. The most precise estimates of the age of this event are based on high-precision radiocarbon dating of different blocks of annual rings from earthquake-killed trees in northern California and southern Washington State. High-precision ages on three different blocks of rings from a stump at a site in northern California limit tree death there to about AD 1700 to 1720. This date is consistent with an age estimate of about AD 1700, obtained by comparing the thickness patterns of the outermost rings of earthquake-killed and old, living trees.

Historical tsunami records from Japan may provide an exact age for the last Cascadia earthquake. Magnitude 8 and 9 earthquakes in Cascadia should produce tsunamis that are large enough to cause significant runup in Japan. Japanese geophysicists have found written records of 1 to 2 m high waves that damaged sites along a 1000 km length of the Japanese coast in January 1700. The waves were not the result of a local earthquake and there are no accounts of a great earthquake elsewhere in the Pacific Ocean that conceivably could be the source of the tsunami. The researchers thus argued, by the process of elimination, that the tsunami was produced by a great earthquake off the west coast of North America. After correcting for the travel time to Japan and the time zone difference, they concluded that the earthquake occurred on 26 January 1700 at about 21:00 h local time. This conclusion agrees with oral traditions of coastal native peoples who describe the shaking and tsunami of a large earthquake at night in winter (e.g. *see* account of Loh̲t'a *in* 'Native oral traditions', above).

The sand sheets of the AD 1700 and 1964 tsunamis are the youngest tsunami deposits that have been found on Vancouver Island; in fact, they are the only tsunami deposits less than 500 years old. At Port Alberni, there are nine older sand and gravel layers that may be tsunami deposits; all are younger than about 5000 years old. On the basis of these data, the average tsunami recurrence on Vancouver Island is about 500 years, which is similar to the average recurrence derived from buried soils in southwestern Washington State. This suggests that most tsunamis large enough to leave a geological signature on Vancouver Island are products of large, local subduction earthquakes and of extremely large subduction earthquakes in Alaska (i.e. comparable in size to the 1964 Alaska earthquake). Many lesser distant earthquakes produced tsunamis, but their runups on the British Columbia coast were smaller than in 1964.

The catalogue of large Cascadia subduction zone earthquakes is probably incomplete. In addition to the seven events recognized in southwestern Washington State, other large earthquakes, which are unrecorded in this area, may have occurred on northern and southern segments of the subduction zone. Conversely, some of the southern Washington State events may not have been large enough to leave geological signatures in, for example, southwestern British Columbia.

HAZARD

Pacific coast

Historical records and geological data show that large, destructive tsunamis strike British Columbia, on average, once every 500 years. These tsunamis are triggered by subduction earthquakes beneath the Pacific Ocean, principally off southern Alaska and along the Cascadia subduction zone. The most vulnerable areas are the outer coast and inlets of Vancouver Island, where damage to some coastal communities would be considerable (Fig. 13). A large Pacific tsunami would attenuate before entering the Strait of Georgia and probably would not cause major damage there.

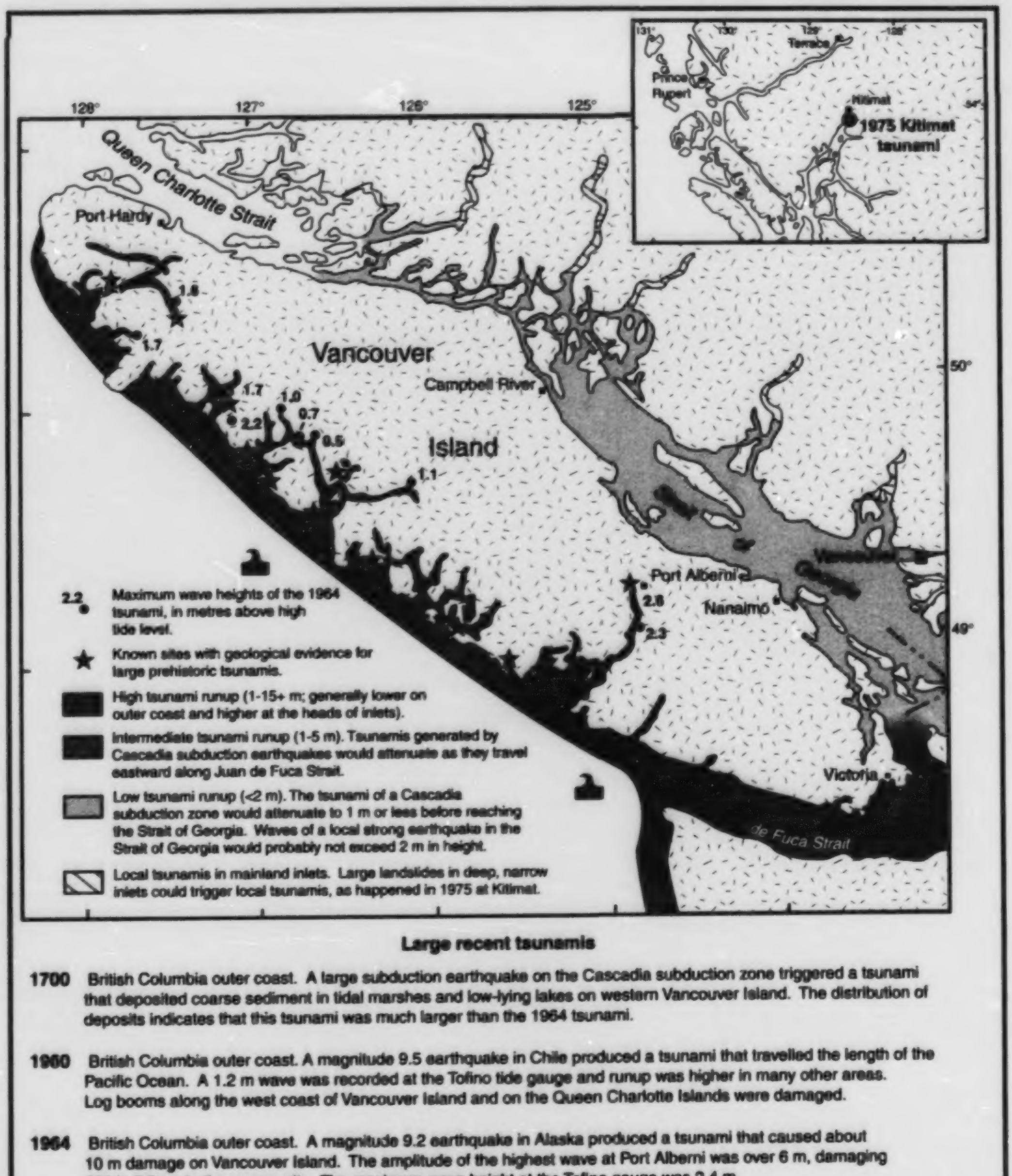
Estimates have been made of maximum water levels and velocities on the British Columbia coast for tsunamis triggered by hypothetical great subduction earthquakes off south-central Alaska, the Aleutian Islands, Kamchatka, and Chile. Researchers have developed computer models to generate and propagate tsunamis from each of these source regions to the British Columbia continental shelf and into 20 inlets. The model predictions were verified against water-level measurements made at tide-gauge stations after the 1964 Alaska earthquake. Simulated seafloor displacements were based on data from the 1960 Chile and 1964 Alaska earthquakes. The model results show that the most vulnerable areas in British Columbia are the west coasts of Vancouver Island and Graham Island, and the central mainland coast. Several sites at the heads of inlets could be hit by waves up to 9 m high, moving at 3 to 4 m/s (Table 2). The largest waves on the northern British Columbia coast were generated from the simulated earthquake off south-central Alaska. In all other regions, the largest waves came from the Aleutian source. Major population centres bordering the Strait of Georgia are not at risk from tsunamis triggered by distant earthquakes because the waves would dissipate as they move through Juan de Fuca Strait and the Gulf Islands.

Table 2. Examples of maximum amplitudes resulting from subduction earthquakes in the North Pacific Ocean.

Sector	Location	Height (m)	Source area
north coast	head of Khutzeymateen Inlet	3.5	Alaska
	Hastings Arm	3–3.5	Alaska
	Stewart	3.5–4	Alaska
north-central coast	Princess Royal Island	3.5–4.5	Shumagin
	Cousins Inlet	4.2	Shumagin
	head of Spiller Channel	7.2	Alaska
	head of Laredo Sound	5.6	Shumagin
	head of Surf Inlet	3.3	Shumagin
south-central coast	heads of Rivers and Moses inlets	3.3–9.2	Shumagin
	Smith Inlet	6–9.3^1	Alaska
northwestern Vancouver Island	Quatsino Sound	5.5–7.2	Shumagin
	Neroutsos Inlet	up to 9	Shumagin
	Quatsino Narrows	8–8.5	Shumagin
	head of Holberg Inlet	3.4	Shumagin
	Forward Inlet	6–7	Shumagin
	Klaskino Inlet	5–6	Shumagin
	head of Quoukinsh Inlet	>10	Shumagin
central western Vancouver Island	Muchalitz Inlet	3.5–4.5	Shumagin
	Muchalat and Tlupana inlets	4.5–>10^1	Shumagin
	south end of Tahsis Inlet	3.5–4.5	Shumagin
	head of Tahsis Inlet	3	Shumagin
	Port Eliza and Espinosa Inlet	3.6–7.6	Shumagin
	Nootka Sound	3.5	Shumagin
southwestern Vancouver Island	Sydney Inlet	3–4	Shumagin
	Pipestem and Effingham inlets	4–8	Shumagin
	Alberni Inlet	3–8^1	Shumagin

Source: Dunbar et al. (1989, Table 12)

[1] Increases towards the head of the inlet.


131°
130°
129°
128°
Terrace
Prince Rupert
Kitimat
54°
1975 Kitimat tsunami
128°
127°
126°
125°
Queen Charlotte Strait
Port Hardy
1.8
1.7
Vancouver
Island
Campbell River
50°
1.7
1.0
0.7
2.2
0.5
1.1
Port Alberni
2.8
2.3
Nanaimo
Strait of Georgia
Vancouver
49°
Victoria
de Fuca Strait
2.2 ● Maximum wave heights of the 1964 tsunami, in metres above high tide level.
★ Known sites with geological evidence for large prehistoric tsunamis.
High tsunami runup (1-15+ m; generally lower on outer coast and higher at the heads of inlets).
Intermediate tsunami runup (1-5 m). Tsunamis generated by Cascadia subduction earthquakes would attenuate as they travel eastward along Juan de Fuca Strait.
Low tsunami runup (<2 m). The tsunami of a Cascadia subduction zone would attenuate to 1 m or less before reaching the Strait of Georgia. Waves of a local strong earthquake in the Strait of Georgia would probably not exceed 2 m in height.
▨ Local tsunamis in mainland inlets. Large landslides in deep, narrow inlets could trigger local tsunamis, as happened in 1975 at Kitimat.

Large recent tsunamis

1700	British Columbia outer coast. A large subduction earthquake on the Cascadia subduction zone triggered a tsunami that deposited coarse sediment in tidal marshes and low-lying lakes on western Vancouver Island. The distribution of deposits indicates that this tsunami was much larger than the 1964 tsunami.
1960	British Columbia outer coast. A magnitude 9.5 earthquake in Chile produced a tsunami that travelled the length of the Pacific Ocean. A 1.2 m wave was recorded at the Tofino tide gauge and runup was higher in many other areas. Log booms along the west coast of Vancouver Island and on the Queen Charlotte Islands were damaged.
1964	British Columbia outer coast. A magnitude 9.2 earthquake in Alaska produced a tsunami that caused about 10 m damage on Vancouver Island. The amplitude of the highest wave at Port Alberni was over 6 m, damaging 260 buildings in the community. The maximum wave height at the Tofino gauge was 2.4 m.
1975	Kitimat. A large submarine landslide near the head of Kitimat Arm in April 1975 generated sea waves that caused $600 000 in damage to wharves and other shore facilities at Kitimat.

Figure 13. *Tsunami hazard map, southwestern British Columbia. The map shows four generalized hazard zones, as well as sites with evidence for large prehistoric tsunamis, the location of the 1975 Kitimat tsunami (inset), and maximum wave heights of the 1964 Alaska tsunami. Details on four important recent tsunamis are given at the bottom (for further information, see Historical tsunamis in Canada in 'Suggested reading'). Southwestern British Columbia is the only part of the Canadian coast where sufficient information exists to produce a map of this type.*

Even larger waves might be produced by an earthquake on the Cascadia subduction zone. A numerical model was used to simulate three tsunamis, involving rupture of different parts of the plate boundary. As might be expected, the west coast of Vancouver Island was found to be the most strongly affected area in British Columbia. For a simulated M8.5 earthquake on the northern section of the subduction zone, maximum wave amplitudes on the outer coast are of the order of 10 m (peak-to-trough height). Amplification, up to a factor of three, is predicted for some inlets (Fig. 14). Much energy is lost as the waves pass through narrow passages connecting Juan de Fuca Strait to Puget Sound and the Strait of Georgia, and tsunami amplitudes are reduced to 1 m or less by the time the waves reach Seattle and Vancouver (Fig. 14). A tsunami would, of course, be superimposed on the tides, which in south-coastal British Columbia have a range of up to 5 m. A positive sea-level displacement at low tide would have much less impact than one at high tide. However, a tsunami could persist through a tidal cycle, thus extreme water levels are likely to combine tsunami and tidal effects.

The travel time of a tsunami with a source on the Cascadia subduction zone is much less than one generated elsewhere in the Pacific Ocean. The first wave would reach the outer coast of Vancouver Island in less than one hour (in fact, in a matter of minutes if the earthquake were centred on the northern part of the subduction zone). Travel times to Victoria and Vancouver might be as little as 1.5 and 3 hours, respectively. In contrast, the 1964 Alaska tsunami arrived on the west coast of Vancouver Island about 4 hours after the earthquake.

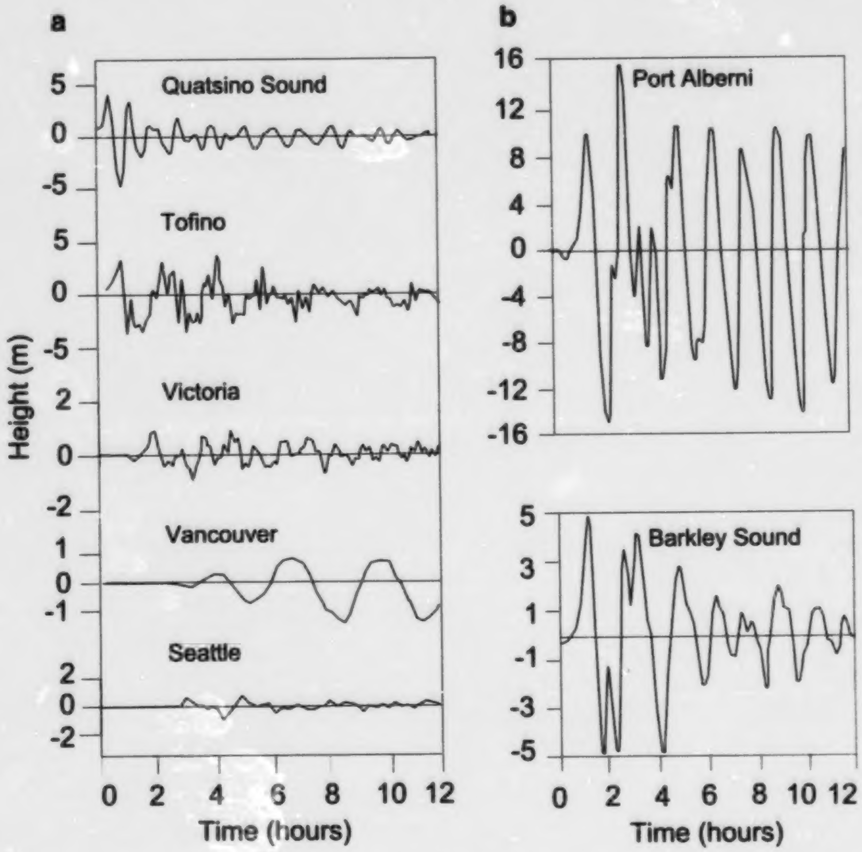


Figure 14. *Sea-level displacements generated by a hypothetical, M8.5 earthquake on the northern Cascadia subduction zone.* ***a)*** *Vancouver, Seattle, and selected sites on Vancouver Island;* ***b)*** *Port Alberni and Barkley Sound. The first wave reaches the west coast of Vancouver Island about 0.5 hour after the earthquake. Much smaller waves arrive at Vancouver and Seattle almost 4 hours later. Resonance in Alberni Inlet is responsible for the approximately threefold amplification of waves between Barkley Sound and Port Alberni* (from *Clague, 1996, Fig. 83;* modified from *Ng et al., 1990).*

Simulations have also been made of tsunamis from large earthquakes within the North American Plate. The earthquakes used in the simulations were similar in magnitude and motion to the 1946 Vancouver Island earthquake, but were centred in Juan de Fuca Strait near Victoria, in Puget Sound near Seattle, and in the Strait of Georgia near Vancouver. In all cases, the resulting tsunamis are small (largest waves = 0.5–1 m) and would have only a local effect (Fig. 15). Little wave energy passes through the Gulf Islands between Juan de Fuca Strait and the Strait of Georgia. It should be noted, however, that a local crustal earthquake of magnitude 8, which is much larger than the 1946 event, might produce a tsunami with waves 3 m or more in height in the Strait of Georgia. A tsunami of this size probably would cause widespread flooding in low-lying areas and might severely damage development along the shoreline. However, no geological evidence has yet been found for such a tsunami.

Although numerical modelling can provide estimates of tsunami amplitudes and velocities, these estimates have large uncertainties. Wave amplitudes are strongly influenced by coastal topography and offshore bathymetry, both of which can be exceedingly complex. As a consequence, tsunami runup can vary by a factor of 5 or 10 over even short stretches of coast.

A landslide triggered by an earthquake or by some other phenomenon could cause a tsunami in one of the fiords on the west coast. The 1964 Alaska earthquake triggered a large (about 75 x 10^6 m^3) submarine slump near Valdez, Alaska, which produced a local tsunami that destroyed waterfront facilities and the fishing fleet (Fig. 2). The slide and accompanying tsunami were responsible for the loss of 30 lives, nearly 25% of all the casualties of the earthquake.

Local landslide-generated tsunamis, unlike tsunamis produced by displacements of the seafloor during earthquakes, affect relatively small areas, although the near-field waves can be very large. A large, rapid landslide is required to generate a significant tsunami. The landslide could be entirely submarine or it could enter the sea from a source on land. The failure could occur in bedrock or in thick Quaternary sediments, involving for example a fiord-head delta.

Atlantic coast

Tsunamis are less of a hazard to the Atlantic coast of Canada than the Pacific coast. Large earthquakes are rare off the east coast of Canada — during historical time only one, the Grand Banks earthquake of 1929, has exceeded magnitude 7. The recurrence interval for an earthquake of the size of the 1929 event, based on the historical record of seismicity, is between a few hundred and one thousand years; however, geological evidence from the Atlantic continental slope and abyssal plain suggests that M7 events may be even rarer than this. Even if an earthquake of this size were to occur off Canada's east cost, it might not directly trigger a tsunami unless it vertically displaced a large area of the seafloor. There is a greater hazard that such an earthquake could indirectly generate a destructive tsunami by triggering a submarine landslide, as happened in 1929. It is also possible that a large, tsunami-generating landslide could occur independently of an earthquake, although there is no historical precedent for such an event.

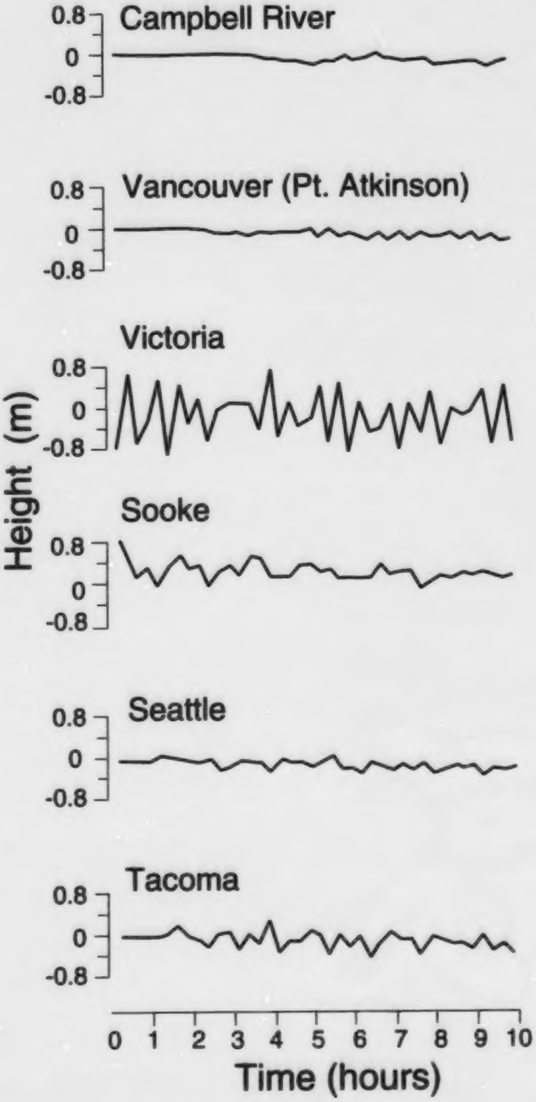


Figure 15. Sea-level displacements generated by a hypothetical, M7, crustal earthquake similar to the 1946 Vancouver Island earthquake, but centred in Juan de Fuca Strait near Victoria. The tsunami is small and localized (from Clague, 1996, Fig. 84; modified from *Murty and Hebenstreit, 1989).*

Arctic coast

Little is known about the tsunami hazard in Arctic Canada. No significant tsunamis have been reported in this region and no geological evidence has been found for prehistoric tsunamis. Although moderate and large earthquakes are possible in some parts of the Arctic, for example Baffin Bay and Beaufort Sea, the presence of extensive sea ice greatly reduces the possibility of a large tsunami anywhere in the region.

HAZARD REDUCTION

Tsunamis cannot be prevented, but the damage they cause can be reduced through a variety of actions. There are two main categories of hazard reduction measures: nonstructural measures, including land-use controls (zoning, relocation, property acquisition), emergency preparedness, and public education; and structural measures, which include dyking, barrier construction, floodproofing, and tsunami-resistant construction.

The first objective of any hazard reduction program is the protection of life. This may require that certain uses of land in high-hazard areas be restricted through zoning regulations or by relocating property owners to higher ground; however, as a rule, these types of action are strongly resisted by residents who would be affected by them. Protection of life also requires that there be an effective system in place to warn people in areas that might be inundated of an approaching tsunami and to provide timely evacuation. An existing tsunami warning system, headquartered in Honolulu and operated by the United States government, provides alerts of possible and actual tsunami dangers following large Pacific earthquakes. The travel times of far-field tsunamis are sufficiently long that threatened low-lying areas in British Columbia could be evacuated following such alerts (Fig. 5). Local tsunamis, however, would strike the British Columbia coast soon after the shaking stopped and, consequently, there would be little time to respond to the emergency. In this context, public awareness of the nature of tsunamis, the chance of one occurring, areas likely to be affected, and the correct response to tsunami warnings are important. A public education program should provide tsunami information at regular intervals, perhaps annually, and should include instructions on how to get information during an alert, where to go, and what things to take.

Structural hazard-reduction measures provide protection from tsunamis either by preventing them from inundating low-lying areas or by lessening their impact on buildings. Dykes can be constructed to prevent waves from reaching threatened residential and commercial areas, although these are typically expensive and should be built to the highest possible level that can be reached by a tsunami. In some cases, offshore barriers can deflect tsunami waves or lessen their energy before they reach the shore. Again, these generally are expensive structures and may provide only limited protection. Elevation of buildings and other types of floodproofing (e.g. installation of seals for basement windows, bolting houses to their foundations) provide protection where water depths are 1 m or less. Structures can be elevated to higher levels to provide greater protection, but costs may be prohibitive.

ACKNOWLEDGMENTS

Thane Anderson and Alan Ruffman kindly provided two of the photographs reproduced here. Tracey Barry drafted the figures. Greg Brooks and Ian Hutchinson reviewed the draft of the text and contributed to its improvement.

REFERENCES

Arima, E.Y., St. Claire, E., Clamhouse, L., Edgar, J., Jones, C., and Thomas, J.
1991: Between Ports Alberni and Renfrew: notes on west coast peoples; Canadian Museum of Civilization, Canadian Ethnology Service, Mercury Series Paper 121, 323 p.

Clague, J.J.
1996: Paleoseismicity and seismic hazards, southwestern British Columbia; Geological Survey of Canada, Bulletin 494, 88 p.

Dunbar, D., LeBlond, P.H., and Murty, T.S.
1989: Maximum tsunami amplitudes and associated currents on the coast of British Columbia; Science of Tsunami Hazards, v. 7, p. 3–44.

Heaton, T.H. and Snavely, P.D., Jr.
1985: Possible tsunami along the northwestern coast of the United States inferred from Indian traditions; Seismological Society of America, Bulletin, v. 75, p. 1455–1460.

Murty, T.S. and Hebenstreit, G.T.
1989: Tsunami amplitudes from local earthquakes in the Pacific Northwest region of North America, Part 2: Strait of Georgia, Juan de Fuca Strait, and Puget Sound; Marine Geodesy, v. 13, p. 189–209.

Ng, M.K.-F., LeBlond, P.H., and Murty, T.S.
1990: Numerical simulation of tsunami amplitudes on the coast of British Columbia due to local earthquakes; Science of Tsunami Hazards, v. 8, p. 97–127.

Whelan, M.
1994: The night the sea smashed Lord's Cove; Canadian Geographic, v. 114, no. 6, p. 70–73.

Wigen, S.O.
1983: Historical studies of tsunamis at Tofino, Canada; *in* Tsunamis — Their Science and Engineering, (ed.) K. Iida and T. Iwasaki; Terra Scientific Publishing Company, Tokyo, Japan, p. 105–119.

Wigen, S.O. and White, W.R.
1964: Tsunami of March 27–29, 1964, west coast of Canada; Canada Department of Mines and Technical Surveys, Ottawa, Ontario, 6 p.

SUGGESTED READING

Historical tsunamis in Canada

Benson, B.E.
1996: The stratigraphy and neotectonic significance of tsunami deposits beneath tidal marshes on Vancouver Island, British Columbia; M.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 112 p.

Clague, J.J., Bobrowsky, P.T., and Hamilton, T.S.
1994: A sand sheet deposited by the 1964 Alaska tsunami at Port Alberni, British Columbia; Estuarine, Coastal and Shelf Science, v. 38, p. 413–421.

Heezen, B.C. and Ewing, M.
1952: Turbidity currents and submarine slumps and the 1929 Grand Banks, Newfoundland earthquake; American Journal of Science, v. 250, p. 849–873.

Iida, K.

1983: Some remarks on the occurrence of tsunamigenic earthquakes around the Pacific; *in* Tsunamis — Their Science and Engineering, (ed.) K. Iida and T. Iwasaki; Terra Scientific Publishing Company, Tokyo, Japan, p. 61–76.

Lockridge, P.A. and Smith, R.H.

1984: Tsunamis in the Pacific basin, 1900–1983; National Geophysical Data Center and World Data Center A for Solid Geophysics, Boulder, Colorado, scale 1:17 000 000.

Murty, T.S.

1977: Seismic sea waves — tsunamis; Fisheries Research Board of Canada, Bulletin 198, 337 p.

1979: Submarine slide-generated water waves in Kitimat Inlet, British Columbia; Journal of Geophysical Research, v. 84, p. 7777–7779.

Murty, T.S. and Boilard, L.

1970: The tsunami in Alberni Inlet caused by the Alaska earthquake of March 1964; *in* Tsunamis in the Pacific Ocean, (ed.) W.M. Mansfield; Proceedings, International Symposium on Tsunamis and Tsunami Research (Honolulu, 1969); East-West Center Press, Honolulu, Hawaii, p. 165–187.

Piper, D.J.W., Shor, A.N., and Clarke, J.E.H.

1988: The 1929 'Grand Banks' earthquake, slump, and turbidity current; *in* Sedimentologic Consequences of Convulsive Geologic Events, (ed.) H.E. Clifton; Geological Society of America, Special Paper 229, p. 77–92.

Thomson, R.E.

1981: Oceanography of the British Columbia coast; Canada Department of Fisheries and Oceans, Canadian Special Publication of Fisheries and Aquatic Sciences, no. 56, 291 p.

Prehistoric tsunamis and earthquakes in Canada

Atwater, B.F., Nelson, A.R., Clague, J.J., Carver, G.A., Yamaguchi, D.K., Bobrowsky, P.T., Bourgeois, J., Darienzo, M.E., Grant, W.C., Hemphill-Haley, E., Kelsey, H.M., Jacoby, G.C., Nishenko, S.P., Palmer, S.P., Peterson, C.D., and Reinhart, M.A.

1995: Summary of coastal geologic evidence for past great earthquakes at the Cascadia subduction zone; Earthquake Spectra, v. 11, p. 1–18.

Benson, B.E., Grimm, K.A., and Clague, J.J.

1997: Tsunami deposits beneath tidal marshes on northwestern Vancouver Island, British Columbia; Quaternary Research, v. 48, p. 192–204.

Clague, J.J. and Bobrowsky, P.T.

1994: Evidence for a large earthquake and tsunami 100–400 years ago on western Vancouver Island, British Columbia; Quaternary Research, v. 41, p. 176–184.

1994: Tsunami deposits beneath tidal marshes on Vancouver Island, British Columbia; Geological Society of America Bulletin, v. 106, p. 1293–1303.

Hutchinson, I., Clague, J.J., and Mathewes, R.W.

1996: Reconstructing the tsunami record on an emerging coast: a case study of Kanim Lake, Vancouver Island, British Columbia, Canada; Journal of Coastal Research, v. 13, p. 545–553.

Satake, K., Shimazaki, K., Tsuji, Y., and Ueda, K.

1996: Time and size of a giant earthquake in Cascadia inferred from Japanese tsunami records of January 1700; Nature, v. 378, p. 246–249.

Native oral accounts and archeological evidence of tsunamis and earthquakes

Clague, J.J.

1995: Early historical and ethnographical accounts of large earthquakes and tsunamis on western Vancouver Island, British Columbia; *in* Current Research 1995-A; Geological Survey of Canada, p. 47–50.

Hutchinson, I. and McMillan, A.D.

1997: Archaeological evidence for village abandonment associated with late Holocene earthquakes at the northern Cascadia subduction zone; Quaternary Research, v. 48, p. 79–87.

Sproat, G.M.

1987: The Nootka; Scenes and Studies of Savage Life [edited and annotated by C. Lillard]; Sono Nis Press, Victoria, British Columbia, 215 p. (originally published as Scenes and Studies of Savage Life; Smith, Elder, London, 1868).

Swan, J.G.

1868: The Indians of Cape Flattery, at the entrance to the straight of Juan de Fuca, Washington Territory; Smithsonian Contributions to Knowledge, no. 220, 108 p.

Predictions of tsunamis amplitudes

Dunbar, D., LeBlond, P.H., and Murty, T.S.

1991: Evaluation of tsunamis amplitudes for the Pacific coast of Canada; Progress in Oceanography, v. 26, p. 115–177.

Hebenstreit, G.T. and Murty, T.S.

1989: Tsunami amplitudes from local earthquakes in the Pacific Northwest region of North America, Part 1: The outer coast; Marine Geodesy, v. 13, p. 101–146.

Murty, T.S.

1992: Tsunami threat to the British Columbia coast; *in* Geotechnique and Natural Hazards; BiTech Publishers, Vancouver, British Columbia, p. 81–89.

Ng, M.K.-F., LeBlond, P.H., and Murty, T.S.

1991: Simulation of tsunamis from great earthquakes on the Cascadia subduction zone; Science, v. 250, p. 1248–1251.

Whitmore, P.M.

1993: Expected tsunami amplitudes and currents along the North American coast for Cascadia subduction zone earthquakes; Natural Hazards, v. 8, p. 59–73.

Landslides

Stephen G. Evans[1]

Evans, S.G., 2001: Landslides; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 43–79.*

Abstract: Canada has a wide range of landslides reflecting the diverse geological and geomorphological environments in the nation's landscape. Rock avalanches and rockfalls are extremely rapid landslides that are common in mountainous regions. Debris avalanches and channelized debris flows, typically triggered by heavy rains, form another group of rapid landslides. Rapid slides and earthflows (flowslides) in Quaternary sediments include landslides in leda clay of the St. Lawrence Lowland. Slumps and earthflows are slow landslides. Landslides occur in permafrost terrain and are sensitive to environmental change. Geotechnical failures in waste dumps, embankments, and excavated slopes form an important group of damaging landslides. Underwater slides are common in Canada's lakes and ocean margins. The secondary effects of landslides (landslide dams and landslide-generated displacement waves) extend the impact of a landslide beyond its debris. Landslides have caused significant loss of life in Canada and have had impacts on Canada's infrastructure, environment, and resources.

Résumé : Il se produit au Canada un vaste éventail de glissements de terrain qui traduisent la variété des contextes géologiques et géomorphologiques du paysage canadien. Les avalanches de pierres et les écroulements constituent des glissements de terrain extrêmement rapides qui sont courants dans les régions montagneuses. Les avalanches de débris et les coulées de débris canalisées, qui sont généralement déclenchées par des pluies intenses, sont un autre groupe de glissements de terrain à déclenchement rapide. Les glissements rapides et les coulées de fluage (glissements par liquéfaction) dans les sédiments du Quaternaire comprennent notamment les glissements dans les argiles à Leda présentes dans les basses terres du Saint-Laurent. Les effondrements et les coulées de fluage sont des glissements de terrain lents. Les glissements de terrain se produisent dans des terrains pergélisolés et sont sensibles aux modifications environnementales. Les glissements géotechniques dans les terrils, les levées de terre et les pentes excavées forment un groupe important de glissements destructeurs. Les glissements sous-marins sont fréquents dans les lacs et les marges océaniques du Canada. Les effets secondaires des glissements de terrain (barrages et vagues de déplacement engendrés par un glissement de terrain) prolongent l'impact d'un glissement de terrain au-delà des débris qu'ils entraînent sur leur passage. Les glissements de terrain sont la cause de nombreuses pertes de vie au Canada et ont des incidences sur les infrastructures, l'environnement et les ressources du pays.

---

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

INTRODUCTION

Landslides are mass movements of geological materials downslope. They may range in size from an individual boulder to large rock masses that form a mountainside, and involve materials ranging from silty clay to massively bedded sedimentary rock. Landslides can be extremely rapid or very slow and occur in a wide variety of geological environments, including under water. Landslides also display movement modes ranging from sliding between relatively intact masses of rock or soil to the flow of completely disintegrated materials. (In this contribution, 'soil' is used in the sense of geotechnical soil, i.e. any geological materials that are not coherent rock. In this context, 'soil' should be not confused with agricultural or pedological soil, i.e. the organic-rich layer that commonly forms the immediate ground cover in Canada. Geotechnical soil includes all types of Quaternary sediments (gravel, sand, silt, and clay), colluvium, as well as earthfill or rockfill materials used to construct embankments. The term may also include various types of soft, soil-like rock, e.g. pyroclastic material. Typically, in contrast to rock, geotechnical soil has no tensile strength except when it is frozen.) In some landslides, the initial movement mode can be transformed into a second or even a third mode, depending on the degree of fragmentation and saturation to which the initial mass is subject. A landslide may also incorporate material along its path that has characteristics different from those of the source material.

Complexities of movement style and materials involved create significant difficulties in landslide classification. Here, landslides are differentiated mainly on the basis of their speed; rapid landslides have mean velocities of 5m/s (18 km/h) or greater, whereas slow landslides move at lower speeds (Fig. 1).

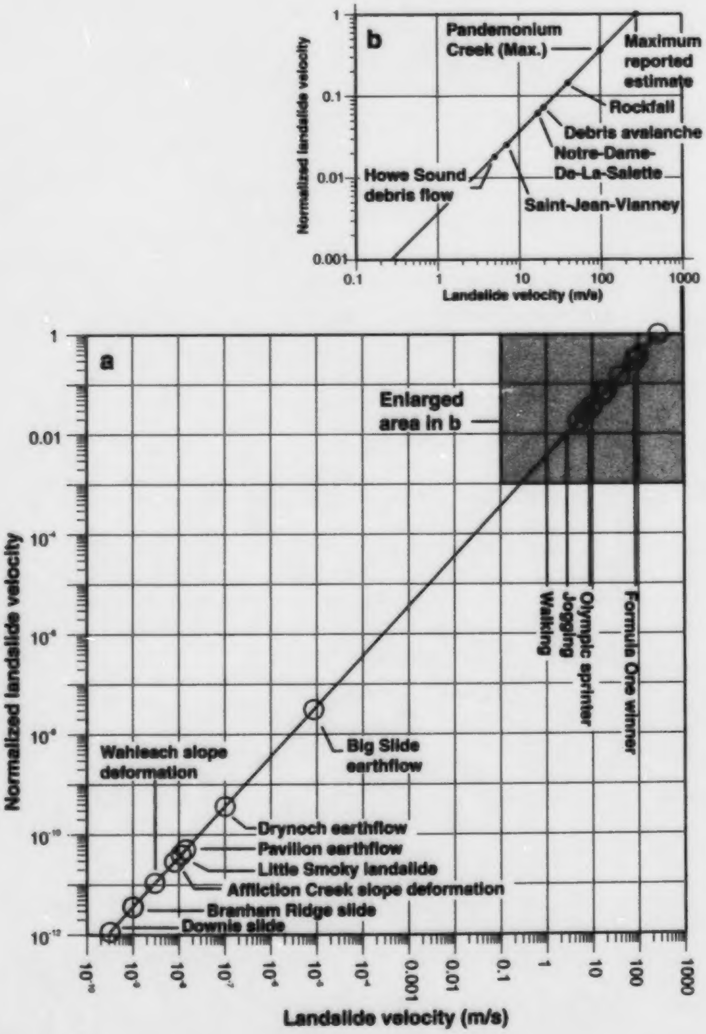


Figure 1.

Log-log plot of the velocities of typical Canadian landslides against the normalized landslide velocity expressed as a ratio of the maximum landslide velocity recorded in the literature (277 m/s or 997 km/h), which was back-calculated from the ballistic trajectory of huge boulders at the 1970 rock avalanche at Huascaran, Peru (Plafker and Ericksen, 1978). Also indicated are the velocities associated with some common recreational activities and major sports.

The wide range of landslide types that occur in Canada reflects the diverse geomorphic and geological environments in the nation's landscape (Fig. 2a). In addition to being an important geomorphic process in many regions of the country, landslides are an important geological hazard and have resulted in considerable property damage and significant loss of life in the historical period, taken to be 1840 to 1998. The hazard posed by landslides is attributable to the physical impact of rapidly moving debris, the movement of ground directly beneath a structure, or the impact of secondary effects such as river damming or landslide-generated waves. Hundreds of millions of dollars have been spent in the successful mitigation of landslide problems in Canada, resulting in a reduction of landslide hazard in many areas.

This contribution provides an overview of major landslide types in Canada with an emphasis on the more destructive landslide types. A review of landslide disasters since 1840 gives an indication of the impact of landslides on Canadian society, and a concluding discussion focuses on the interaction between landslides and the environment. Localities mentioned in the text are shown in Figure 2b.

RAPID LANDSLIDES IN ROCK SLOPES

Rock avalanches

Rock avalanches involve the initial failure and subsequent disintegration of a large rock mass on a mountain slope and the rapid downslope movement of this debris into a valley or onto a glacier surface (Fig. 3). The term was first used by Geological Survey of Canada geologists to describe the 1903 Frank Slide, in the Rocky Mountain Foothills of Alberta (Fig. 2b, 4). Rock avalanches typically involve volumes greater than 10^6 m^3 and can be very mobile (Fig. 5), i.e. debris may travel long distances (up to 10 km in some examples from the Cordillera) from their source. When channelled in deep, narrow valleys, the debris may exhibit sharp changes in direction. As the debris passes through bends, its surface superelevates on the outside of the bends, forming trimlines that are higher than the inside trimline (Fig. 5a). Debris may also run up to considerable heights on opposing slopes (Fig. 5a, b) and may contain particles ranging in size from fine sand to massive, house-sized boulders (Fig. 6).

Rock avalanches are the fastest type of landslide and may reach very high speeds. According to eyewitnesses, for example, the 1903 Frank rock avalanche (Fig. 4) travelled over 3.12 km in about 100 seconds, an average velocity of 31.2 m/s (112 km/h). In the 1959 Pandemonium Creek rock avalanche in the Coast Mountains of British Columbia (Fig. 2b, 5a), the debris may have reached speeds of up to 100 m/s (360 km/h; Fig. 1). Some rock avalanches have generated winds strong enough to blow down trees adjacent to the landslide path. Rock avalanches are very destructive when they affect human activity as exemplified by the 1903 Frank Slide (Fig. 4), which struck the outskirts of the town of Frank, Alberta, killing about 75 people.

In Canada, rock avalanches are most common in the Cordillera where hundreds of prehistoric rock-avalanche deposits exist, the largest being the Valley of the Rocks deposit (estimated volume 10^9 m^3) in the Rocky Mountains of southeastern British Columbia (Fig. 2b). Twenty-nine high-velocity rock avalanches are known to have occurred in the Cordillera since 1855, the largest being the 1965 Hope Slide (estimated volume 48 x 10^6 m^3; Fig. 2b, *see also* cover illustration). Rock-avalanche deposits have also been noted in mountainous areas of Canada's Arctic Archipelago. At Ptarmigan Fiord on Baffin Island (Fig. 2b), for example, debris from an undated rock avalanche (estimated volume 40 x 10^6 m^3) involving granitic gneiss extends to the opposite shoreline, blocking the fiord. No rock avalanche has been documented in the archipelago in the historical period.

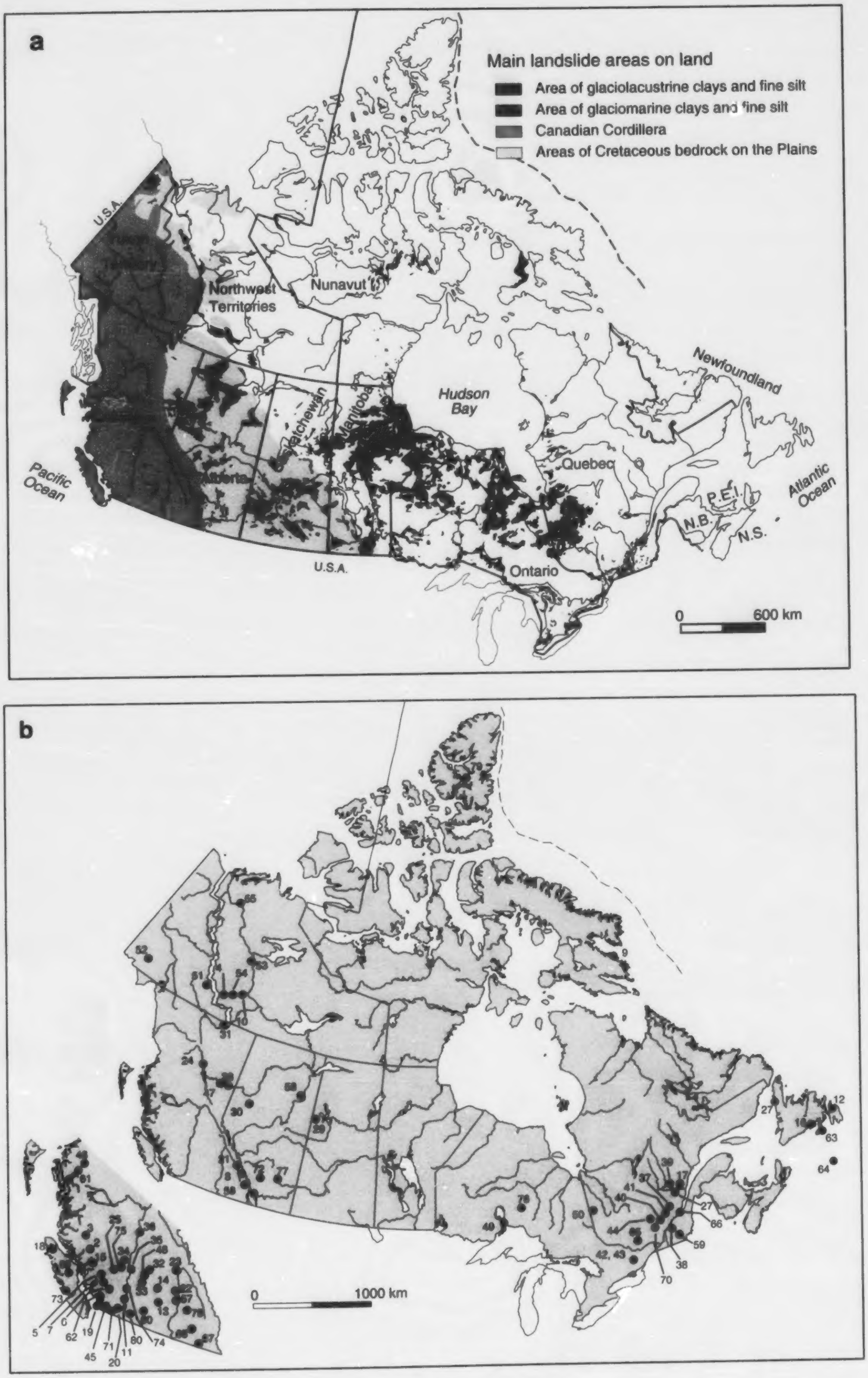
The flanks of volcanoes are particularly prone to catastrophic failure; rock avalanches that originate on the slopes of volcanoes are termed 'volcanic rock avalanches'. (Volcanic rock avalanches are sometimes confusingly referred to as 'debris avalanches'. In this contribution, the latter term is used to refer to smaller scale movements of debris on steepland slopes as discussed in 'Debris avalanches and channelized debris flows'.) Several have taken place in the twentieth century on the Quaternary volcanoes of the Garibaldi Volcanic Belt, southwestern British Columbia, the largest being the 1975 Devastation Glacier rock avalanche (estimated volume 12 x 10^6 m^3; Fig. 2b). Like many other volcanoes in the world, volcanoes of the Garibaldi Volcanic Belt have also been subject to massive, large-scale flank collapse in the Quaternary. The Late Pleistocene collapse of the western flank of Mount Garibaldi (Fig. 2b) involved 2.5 x 10^9 to 3.0 x 10^9 m^3 of material, a volume comparable to that of the 1980 rockslide-avalanche at Mount St. Helens in Washington State. Other massive flank collapses occurred in the Holocene in the Garibaldi Volcanic Belt at Mount Cayley (Fig. 2b) and at the Mount Meager Volcanic Complex (Fig. 2b), the latter event being associated with the explosive eruption of Plinth Peak (Fig. 2b) at about 411 BC. Both events temporarily blocked major rivers in the southern Coast Mountains.

Rapid rockslides

Rapid rockslides are transitional to rock avalanches. They occur when a rock mass slides on a detachment surface, such as a bedding plane or fault surface, and only part of the debris leaves the source area (Fig. 7). Most of the debris is deposited only a limited distance downslope (Fig. 7).

Rockfalls

Rockfalls involve a much smaller rock mass than rock avalanches or rapid rockslides and this rock mass disintegrates into numerous blocks that fall, bounce, and roll on steep slopes after detaching from a rock slope. Rockfalls typically involve less than 50 000 m^3 of material, but may attain high velocities in the range of 35 to 40 m/s (126–144 km/h; Fig. 1). They are a constant problem along transportation routes


a
Main landslide areas on land
Area of glaciolacustrine clays and fine silt
Area of glaciomarine clays and fine silt
Canadian Cordillera
Areas of Cretaceous bedrock on the Plains
U.S.A.
Yukon Territory
Northwest Territories
Nunavut
Hudson Bay
Newfoundland
Quebec
Saskatchewan
Manitoba
Alberta
Pacific Ocean
Atlantic Ocean
P.E.I.
N.B.
N.S.
Ontario
U.S.A.
0
600 km
b
0
1000 km
9
12
27
16
63
64
17
39
37
41
40
27
66
44
65
59
38
42, 43
70
60
49
78
29
58
30
24
31
10
84
53
51
52
55
77
72
5
7
6
62
19
45
71
20
11
80
74
73
18
67
22
32
14
13
23

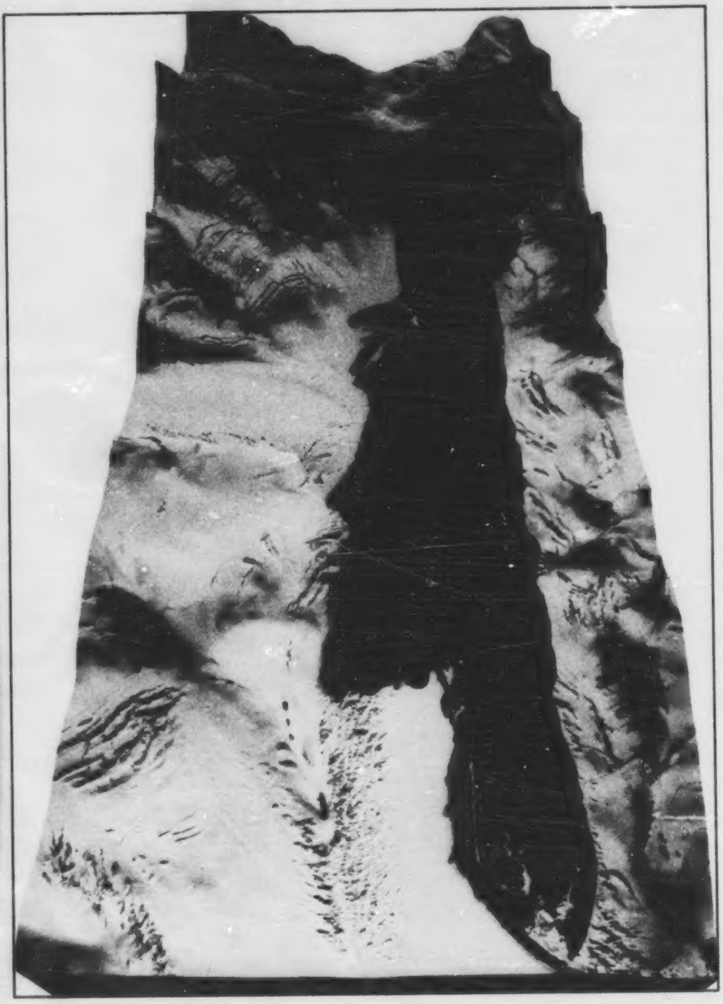


Figure 3.

Perspective view to the southeast of the 1997 Mount Munday rock avalanche on Ice Valley Glacier, Waddington Range, southern Coast Mountains. This digital image was prepared from aerial photographs flown on 20 August 1997 and consists of a digital elevation model with an orthophoto drape. Note flow lines in the debris. The top of the source area is at 3000 m a.s.l. and the lower tip of the debris is at 2100 m a.s.l. The length of the rock-avalanche path is 4.7 km.

Figure 2. a) *Distribution of landslide-prone materials (Cretaceous shale units, Pleistocene glaciolacustrine and glaciomarine sediments) and the limit of the Cordillera where the impact of landslides in Canada is most concentrated.* ***b)*** *Map of Canada showing localities mentioned in the text. 1, Frank Slide; 2, Mount Munday; 3, Pandemonium Creek; 4, Avalanche Lake; 5, Mount Meager Volcanic Complex including Devastation Glacier, Mount Meager, Affliction Creek, and Plinth Peak; 6, Mount Garibaldi; 7, Mount Cayley; 8, Valley of the Rocks; 9, Ptarmigan Fiord; 10, North Nahanni rockslide; 11, Yale, Fraser Canyon; 12, Upper Island Cove, 13, Belgo Creek; 14, Hummingbird Creek; 15, Klattasine Creek; 16, Harbour Breton; 17, La Baie; 18, Port Alice; 19, Howe Sound including Lions Bay, Charles Creek, Woodfibre, and Britannia Beach; 20, Laidlaw; 21, Mount Stephen; 22, Downie Slide; 23, Little Chief; 24, Branham Ridge; 25, Santa Claus Mountain; 26, Mount Éboulé; 27, Bonne Bay; 28, Fort St. John and Peace River bridge; 29, Deer Creek bridge; 30, Little Smoky River landslide; 31, Scatter River; 32, Ashcroft; 33, Drynoch Slide; 34, Pavilion Slide; 35, Big Bar; 36, Big Slide, Quesnel; 37, Saint-Alban; 38, Nicolet; 39, Saint-Jean-Vianney; 40, Saint-Luc-de-Vincennes; 41, Saint-Thuribe; 42, Lemieux; 43, South Nation River; 44, Saint-Boniface-de-Shawinigan; 45, Haney; 46, Terrace (Mink Creek); 47, Attachie; 48, Spences Bridge; 49, Nipigon River; 50, Duparquet; 51, Surprise Rapids; 52, Steele Creek; 53, Old Fort Point; 54, Dekale Creek; 55, Thunder River; 56, Fording River (East Kootenay Coalfield); 57, Revelstoke dam; 58, Syncrude pit; 59, Asbestos; 60, Brenda mine; 61, Kitimat; 62, Fraser River delta; 63, Burin Peninsula; 64, Grand Banks; 65, Notre-Dame-de-la-Salette; 66, Québec; 67, Creston; 68, Kootenay Lake; 69, Mount Colonel Foster; 70, Saint-Maurice River; 71, Calgary; 72, Chilliwack; 73, Clayoquot; 74, Hell's Gate (Fraser Canyon); 75, Lillooet; 76, Nakina; 77, Dorothy; 78, Alkolkolex River; 79, Sawtooth Range; 80, Hope Slide.*

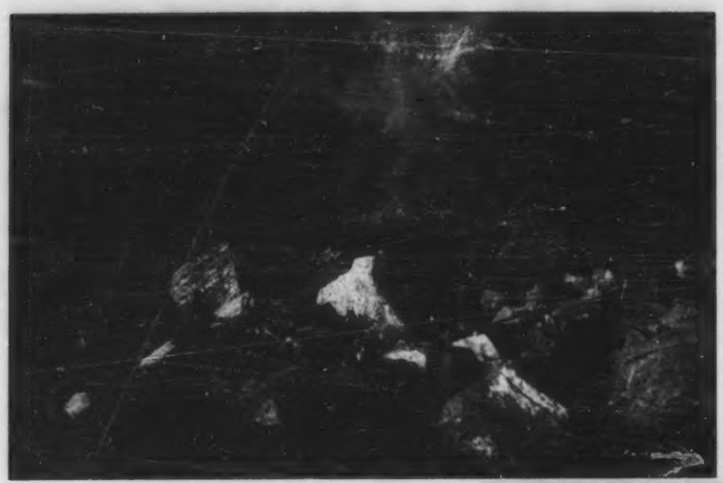


Figure 4.

The 1903 Frank Slide, Alberta. View is from the distal limit of debris towards the detachment zone on Turtle Mountain. Note the characteristics of limestone debris and the person for scale. Photograph by S.G. Evans. GSC 2000-078

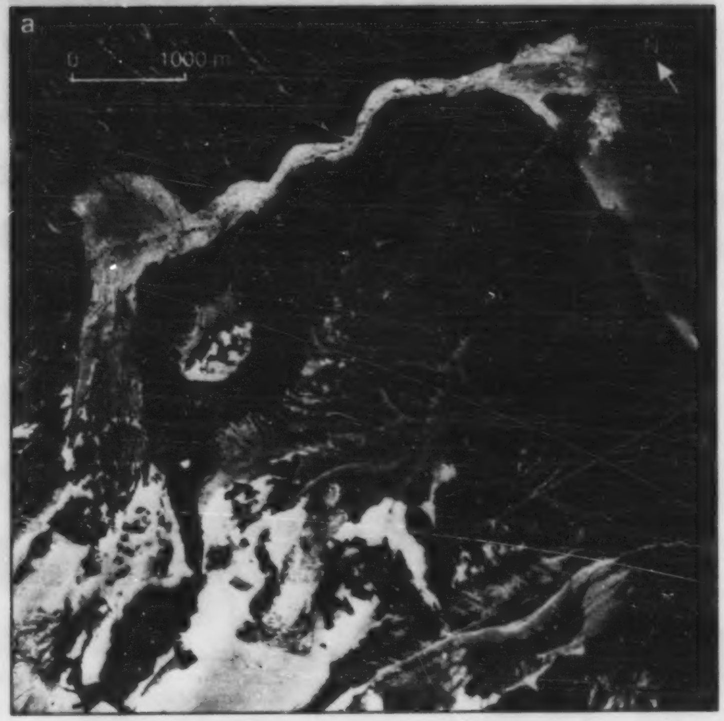
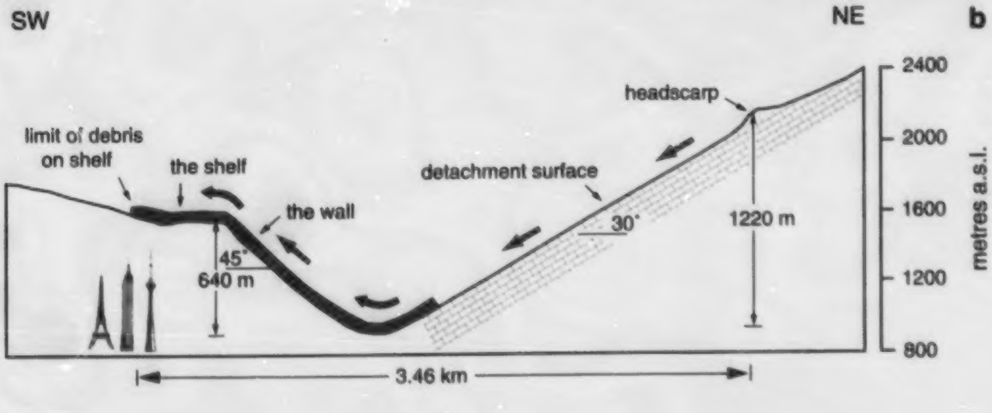


Figure 5.

***a**) Vertical aerial photograph showing the path of the 1959 Pandemonium Creek rock avalanche, Coast Mountains, British Columbia. The volume of the initial failure in granitic gneiss (arrow) was approximately 5 x 10^6 m^3. The total distance travelled was approximately 9 km over a vertical distance of about 2 km. Photograph taken on 31 July 1965. Province of British Columbia BC5145-163 **b**) Extraordinary runup of rock-avalanche debris at the Avalanche Lake rock avalanche, which occurred about AD 1440 in the Mackenzie Mountains, Northwest Territories. The landslide involved 200 x 10^6 m^3 of Devonian dolomite dipping towards the valley. The vertical runup onto the shelf is 640 m. For comparison, the Eiffel Tower (320 m), the Empire State Building (443 m), and Toronto's CN Tower (553.33 m) are shown at the same scale. The runup at Avalanche Lake is the highest recorded in the world.*


SW
NE
b
headscarp
limit of debris on shelf
the shelf
detachment surface
the wall
30°
1220 m
45°
640 m
3.46 km
2400
2000
1600
1200
800
metres a.s.l.

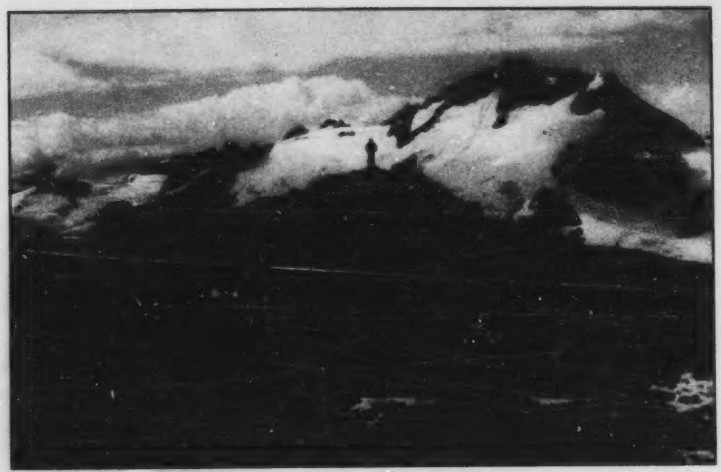


Figure 6.

Debris of the Mount Munday rock avalanche on Ice Valley Glacier, Coast Mountains, British Columbia. Note the wide range of particle sizes typical of rock-avalanche debris. The person is standing on a huge gneissic block approximately 20 m in longest dimension. Photograph by S.G. Evans. GSC 2000-079

Figure 7.

Oblique aerial photograph of the North Nahanni rockslide, Northwest Territories. The landslide was triggered by the October 1985 North Nahanni earthquake (M6.6) and involved 8 x 10^6 m^3 of Devonian limestone. Relief in the vicinity of the rockslide is 476 m. Photograph by S.G. Evans. GSC 2000-080

through rocky terrain across Canada (Fig. 8a) and have also affected buildings located below bedrock slopes (Fig. 8b). A number of deaths have resulted from rockfalls in the historical period. Exposure to rockfall hazard along transportation routes has been reduced by rock-slope stabilization, rock-slope protection, and the installation of rockfall warning fences. In other cases, rockfall hazard has been reduced by the excavation of ditches or the erection of cable net fences across a slope, both of which are designed to catch rolling or bouncing boulders before they reach buildings or transportation routes farther downslope.

DEBRIS AVALANCHES AND CHANNELIZED DEBRIS FLOWS

Debris avalanches and channelized debris flows form a group of landslides that are characterized by the rapid flow of saturated materials in steepland terrain. Typically, debris avalanches and debris flows involve boulder- and gravel-sized clasts that are transported in a finer grained slurry (Fig. 9).

A debris avalanche occurs when a saturated mass fails on an open slope and does not become channelized. It is differentiated from a channelized debris flow, which occurs when a saturated mass of debris moves downslope confined in a steep stream channel.

Debris avalanches are common in forested steepland terrain. Landslide volumes generally range from 20 000 to 60 000 m^3. Debris-avalanche paths may exceed 500 m in length, but involve only a very thin veneer (less than 2 m) of surficial materials. On the basis of eyewitness accounts, debris avalanches commonly travel at over 10 m/s (36 km/h; Fig. 1). An analysis of the 1990 Belgo Creek debris avalanche (estimated volume 23 000 m^3; Fig. 2b, 10a) indicated that local velocities reached as high as 20 m/s (72 km/h). The behaviour and occurrence of debris avalanches contrasts with those of channelized debris flows, which generally recur in a stream channel path and whose velocity rarely exceeds 7 to 8 m/s (25–28 km/h).

Debris flows are frequently generated by the channelization of an initial debris avalanche triggered in the steep upper slopes of a watershed (Fig. 10b). In other cases,

Figure 8.

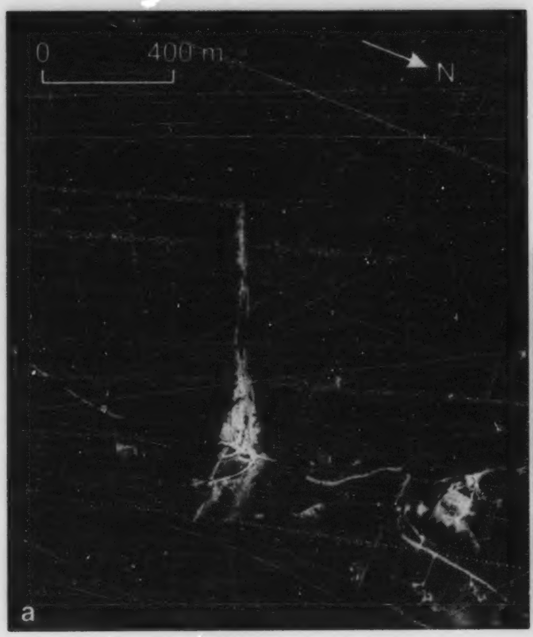
a*) Rockfall fragments on the Trans-Canada Highway, 2 km east of Yale, Fraser Canyon, British Columbia. Note the person for scale. The rockfall was triggered by heavy rains on 9 November 1990. Photograph courtesy of Duncan Wyllie, Golder Associates.* ***b****) Rockfall fragment that damaged buildings and crushed a car at Upper Island Cove, Newfoundland, 14 February 1999. Photograph courtesy of* The Telegram.

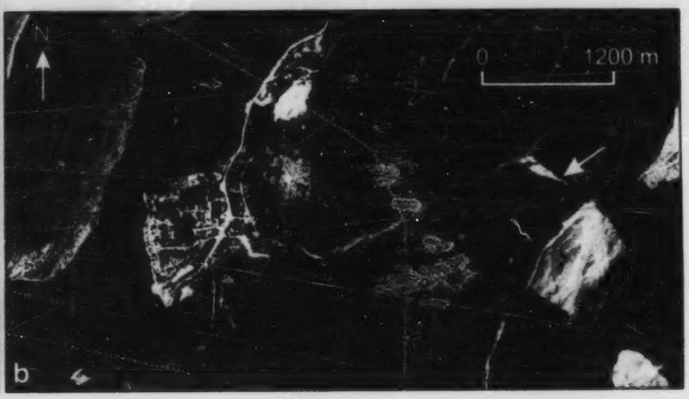
Figure 9.

Debris-flow deposit in the Alkolkolex River valley, Columbia Mountains, British Columbia. Note the characteristic bouldery texture and inverse grading of the deposit (i.e. coarser boulders on the surface). The section was exposed by post-event erosion. Note the person for scale. Photograph by S.G. Evans. GSC 2000-081

Figure 10.

***a**) Vertical aerial photograph of the 1990 Belgo Creek debris-avalanche site, southern British Columbia. Three people were killed when the debris avalanche crushed their home just upslope from the road. Photograph taken on 2 August 1993. Province of British Columbia 30BCC93061;13. **b**) Vertical aerial photograph of the 1997 Hummingbird Creek debris flow, south of Sicamous, British Columbia. The debris flow was initiated as a debris avalanche in the upper part of the drainage basin (arrow) and flowed down the creek bed, damaging houses on the fan at Swansea Point (S) on the shores of Mara Lake. Province of British Columbia 15BCCB97069-33*





stream-bed material may be mobilized when stream discharge exceeds a critical threshold, for example during a severe rainstorm or as a result of the sudden draining of an ice-dammed or moraine-dammed lake. In Klattasine Creek, British Columbia (Fig. 2b), for example, a massive debris flow (estimated volume ~2×10^6 m^3) was triggered in the early 1970s when a moraine-dammed lake burst in the headwaters of the creek. Debris flows may also be triggered by the impact loading of stream-bed materials by rockfalls or falls of glacier ice.

Debris flows and debris avalanches occur widely in the Cordillera of Western Canada. They also occur in the Appalachian and Laurentian mountains of Eastern Canada where they are invariably linked to heavy rains. Two debris avalanches triggered by heavy rains led to a total of six deaths in 1973 at Harbour Breton, Newfoundland (Fig. 2b), and in 1996 at La Baie (Fig. 2b), in the Saguenay region of Quebec, when they struck houses located at the base of the failed slope.

Debris flows are a major problem in southern British Columbia where a number of defensive structures have been constructed to protect sites from their impact. The first debris-flow defence structures built in Canada were deflection dykes constructed at Port Alice on Vancouver Island (Fig. 2b) after destructive debris flows struck the town in 1973 and 1975. The dykes, 3 and 4.5 m high, were placed at the head of a fan to deflect debris flows to the margins of the fan and away from the town (Fig. 11). The protective works cost a total of $250 000 when they were built in 1976.

Since 1976, a number of debris-flow defensive structures have been constructed in the Cordillera of Western Canada. These include debris retention structures that stop and dewater debris in containment basins upstream of debris fans, channelization works that confine the debris in its passage over a fan surface, and deflection berms that either divert the flow away from potential impacts on infrastructural facilities into a predetermined deposition area, or create open containment areas.

The most sophisticated structures have been constructed at several locations along Howe Sound near Lions Bay (Fig. 2b), just north of Vancouver, at a total cost of about $20 million (1986 dollars). At Charles Creek, for example, a large debris retention basin (Fig. 12) was constructed in the mid-1980s to protect transportation routes and homes below. The structure was designed to retain a debris flow event of 29 000 m^3.

Deflection berms have been constructed at several locations to protect major transportation routes from debris flows (Fig. 13a). At Mount Stephen in the Rocky Mountains, a chronic debris flow and snow avalanche problem on the Canadian Pacific Railway was mitigated by the combination of an upslope deflection berm and a concrete debris shed to carry the debris flow over the railway (Fig. 13b). The structures were completed in 1987 and deflected and carried debris flows over the railway in 1988, 1989, and 1994. Less costly deflection structures have been constructed in the Cordillera to protect transmission towers and salmon spawning streams.

Figure 11. Oblique aerial photograph of Port Alice, Vancouver Island, British Columbia, taken in 1983, showing debris deflection dykes (arrow) constructed in 1976 to protect the town from debris flows. Photograph by S.C. Evans. GSC 2000-082

SLOW SLIDES

Slow slides occur in both rock and soil across Canada and are the most common type of landslide in the country. They occur when a mass of soil or rock slowly slides downslope on a discrete shear surface (or defined shear zone). Although disruption of the initial mass takes place during movement (e.g. the formation of slide blocks), total disaggregation does not occur and a part of or all the moving mass remains on the sliding surface. Sliding is frequently initiated by an increase in water pressure on the sliding surface, which reduces the resistance to sliding. The sliding surface may vary in shape from a partial circle to a simple translational plane. However, compound sliding surfaces consisting of a combination of curved and planar elements are most common, reflecting the heterogeneous nature of most geological materials in which failure occurs.

Slides in rock have been reported from many parts of the Cordillera. In southeastern British Columbia, for example, large-scale compound slides involving high mountain slopes are common in the metamorphic rocks of the Columbia Mountains. They include the Downie Slide and the Little Chief Slide (Fig. 2b, 14) located within BC Hydro reservoirs in the Columbia River valley. These landslides have been successfully stabilized by complex underground drainage works consisting of tunnels and rock drains.

Compound slides also occur in gently folded sedimentary rocks in parts of the Rocky Mountains and the Mackenzie Mountains where resistant limestone or sandstone overlies weaker shale. The caprock forms well defined scarps and the sliding surface is seated in the shale. Movement is generally slow. At Branham Ridge in the Rocky Mountains (Fig. 2b) of northern British Columbia, for example, a massive compound slide involves an estimated 2.5×10^9 m^3 of material (Fig. 15). The slide is seated in shale and the weak marine mudrock is overlain by more resistant sandstone that forms the prominent scarp of the landslide. Sliding is taking place in a shear zone parallel to the bedding of the shale, which dips

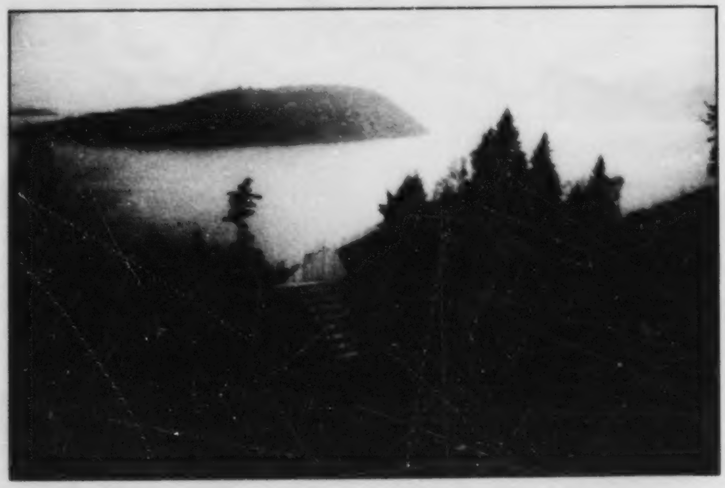


Figure 12.

Debris-flow retention structure on Charles Creek, on the eastern shore of Howe Sound, British Columbia, constructed to protect homes downstream in the mid-1980s. View is downstream. Howe Sound is visible in the background. Photograph by S.G. Evans. GSC 2000-083

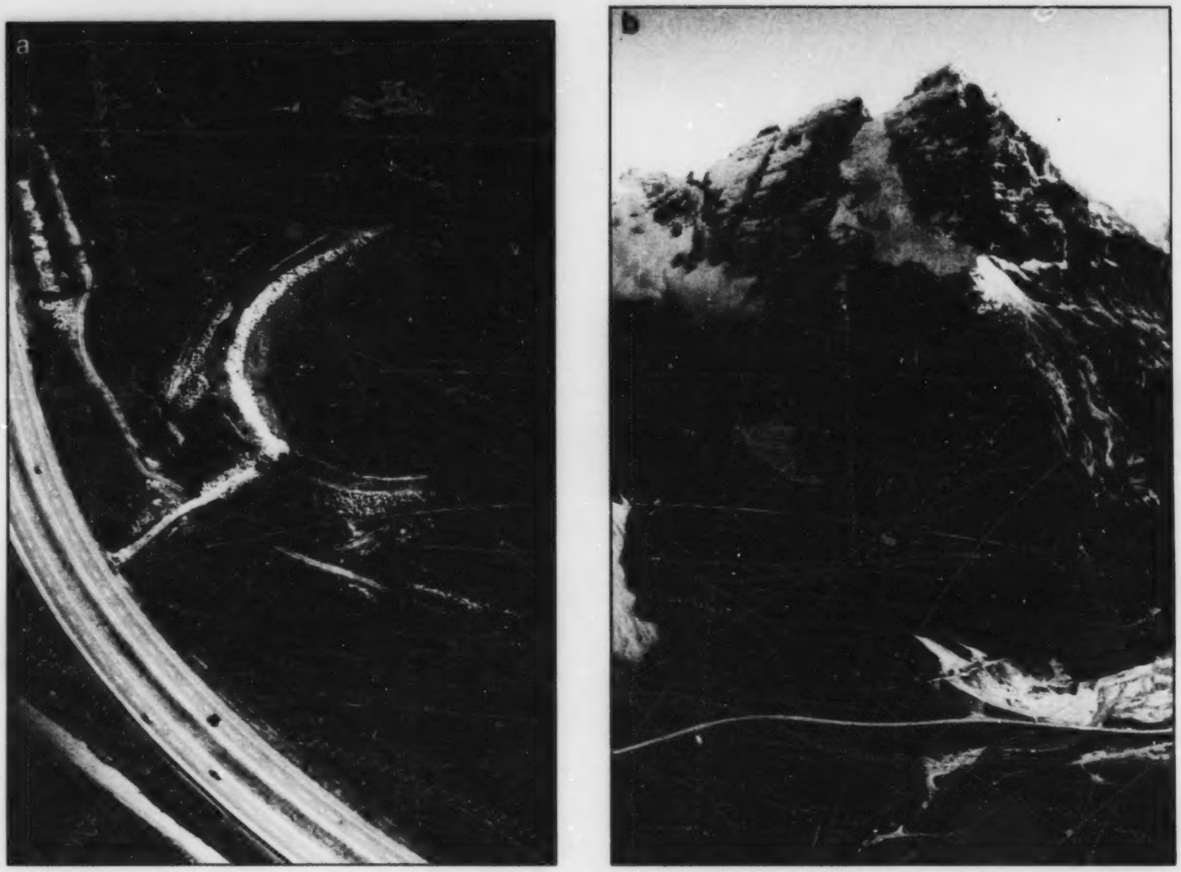


Figure 13. ***a)*** *Debris-flow deflection barrier constructed to protect the Trans-Canada Highway and the Canadian National Railway track near Laidlaw, Fraser River valley, British Columbia. The arrows show the direction of the debris flow. Photograph by S.G. Evans. GSC 2000-084* ***b)*** *Oblique aerial view of Mount Stephen (3185 m a.s.l.), southern Rocky Mountains, British Columbia, and the debris-flow track (partly in shadow) beneath the hanging glacier. The track has been subject to periodic debris-flow and snow-avalanche activity during the twentieth century. Note channelization works above the CP Rail track, which passes under the debris-flow/avalanche shed. These works were constructed in 1987. Note the Trans-Canada Highway and the Kicking Horse River. Photograph by R. Couture, taken in August 1999. GSC 2000-085*

Figure 14.

Oblique aerial photograph of the Downie Slide, a massive rockslide covering 9 km² on the west side of the Columbia River, Columbia Mountains, British Columbia. The slide is located within the reservoir of the Revelstoke dam (Lake Revelstoke), which submerged the toe of the landslide by more than 70 m when it reached full pool level (573 m a.s.l.) in 1984. The cleared areas on the slide are BC Hydro drill sites for subsurface investigations. The landslide has an estimated volume of 1.5 x 10^9 m³ and extends from 502 m a.s.l. at river level to a prominent headscarp at 1520 m a.s.l. The slide involves movement in complex high-grade schist and gneiss; sliding takes place on penetrative foliation that dips about 18° towards Lake Revelstoke, parallel to the topographic slope. Photograph by S.G. Evans. GSC 2000-086

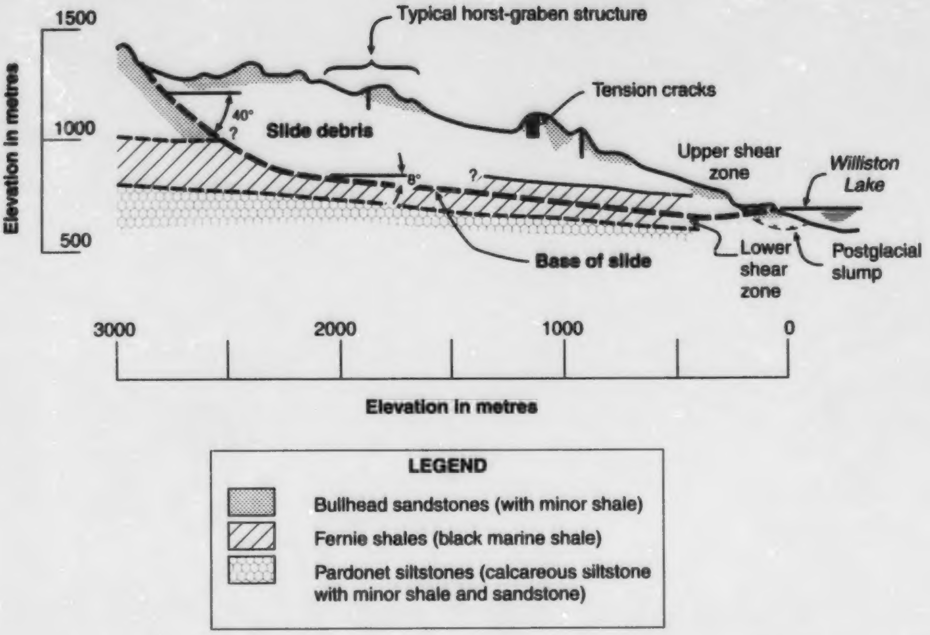

1500
1000
500
Elevation in metres
Typical horst-graben structure
Tension cracks
40°
?
Slide debris
8°
?
Upper shear zone
Williston Lake
Base of slide
Lower shear zone
Postglacial slump
3000
2000
1000
0
Elevation in metres
LEGEND
Bullhead sandstones (with minor shale)
Fernie shales (black marine shale)
Pardonet siltstones (calcareous siltstone with minor shale and sandstone)

Figure 15. *Cross-section of the Branham Ridge landslide, northern Rocky Mountains, British Columbia* (modified from *Enegren and Moore, 1992, Fig. 1*). *The landslide is within the BC Hydro reservoir (Lake Williston) behind the W.A.C. Bennett dam.*

towards the valley at 8°. Horizontal movements of up to 3.5 cm/a (Fig. 1) were measured during the period 1972 to 1990.

Compound slides are especially common in both Tertiary volcanic and sedimentary rocks of British Columbia and the Yukon Territory. In the Interior Plateau of British Columbia, Tertiary lava flows overlie weak, tuffaceous, sedimentary rocks, giving rise to numerous, large-scale landslides.

Mountain-slope deformation is a problematic process related to slow sliding and consists of slow, deep-seated movement of a large rock mass that commonly exhibits loosening and fracturing in the subsurface and signs of disturbance on the slope surface. The process is termed 'sagging' and may represent an early phase in the development of deep-seated landsliding in mountainous terrain. This type of slope movement may involve movement along discrete shear surfaces and/or deep-seated mass creep. It is manifested in topographic features such as cracks, fissures, trenches, antislope scarps (bedrock scarps that face uphill and are formed by differential rock slope deformation) at mid- or upper-slope locations, and, in some cases, slope bulging at lower slope locations (Fig. 16). These linear geomorphic

←

Figure 16. *Oblique aerial photograph of sackung features on Santa Claus Mountain, southern Coast Mountains, British Columbia. Movement has taken place in rocks of the Bridge River Group. Seton Lake is in the background. Photograph by S.G. Evans. GSC 2000-087*

features are termed 'sackungen', after the German word for sagging. They generally occur without well defined headscarps, lateral scarps, or lateral shear zones, suggesting that slope movement is occurring without the formation of well defined shear surfaces, unlike rockslides. Sackungen are widespread throughout the Cordillera; movements associated with them have damaged hydroelectric facilities in southwestern British Columbia. In some cases where scarps are linearly extensive, a tectonic origin (i.e. an earthquake fault) has been suggested.

In Eastern Canada, rockslides and sagging slopes have been documented in the Laurentian Mountains of the Saguenay region (e.g. Mount Éboulé; Fig. 2b), in Newfoundland (e.g. in the Bonne Bay area; Fig. 2b), in the mountains of Labrador, and on Baffin Island.

Slides are widespread in the soft Cretaceous shale units along the eastern margin of the Cordillera and in the Prairie Provinces (Fig. 2b, 17). Sliding in these materials may occur along a curved, curvilinear, or multiplanar surface and is usually retrogressive (the term 'retrogressive' is used to describe the headward progression of a landslide through the development of a series of failures along a succession of failure surfaces as the movement involves an increasing area of terrain; the direction of retrogression is opposite to the movement of the landslide debris) (Fig. 18). Basal shear surfaces in the shale units are often localized in bentonitic interbeds (i.e. clay layers) of very low shear strength. The slides are generally slow-moving, but have damaged or destroyed structures founded on the moving mass. The landslide susceptibility of these rocks has posed particular problems for road, rail, and pipeline crossings of Prairie valleys, which are commonly incised into Cretaceous bedrock (Fig. 19). For example, a slide in Cretaceous shale units in the north abutment destroyed the Alaska Highway suspension bridge over the Peace River near Fort St. John, British Columbia, in October 1957 (Fig. 2b). The cost of dismantling and replacing the bridge was in the order of $60 million (converted to 1986 dollars). This was probably Canada's most costly single landslide in the

Figure 17.

Landslide in Cretaceous shale of the Bearpaw Formation in the Red Deer River valley, near Dorothy, Alberta. The dramatic graben structure, formed by differential vertical movement at the head of the slide, is a result of complex sliding on a horizontal sliding surface (Scott and Brooker, 1968, Plate III). GSC 118541

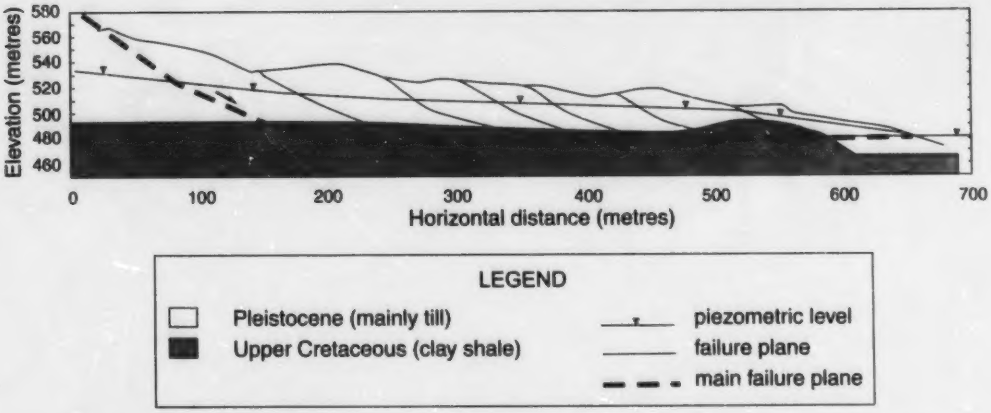


***Figure 18.** Cross-section of a retrogressive landslide involving basal sliding in Cretaceous shale, at the crossing of the Little Smoky River, central Alberta. Note the presence of retrogressive slide blocks* (modified from *Thomson and Hayley, 1975, Fig. 12).*

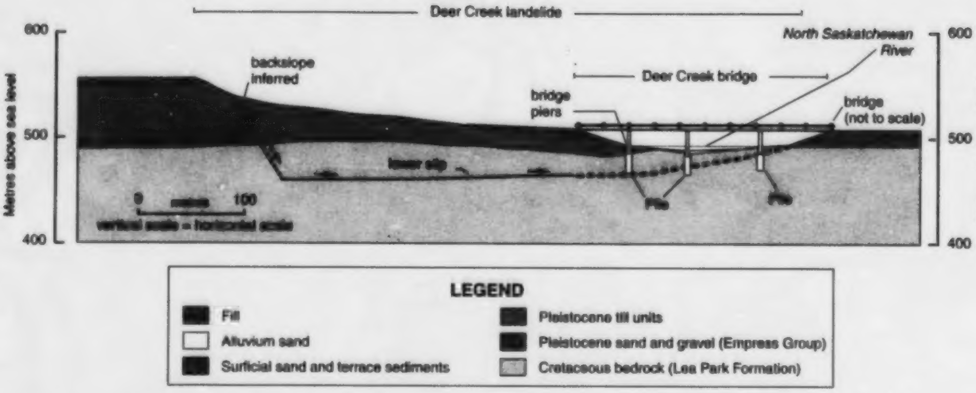

Deer Creek landslide
600
500
400
Metres above sea level
backslope inferred
North Saskatchewan River
Deer Creek bridge
bridge piers
bridge (not to scale)
lower slip
0 metres 100
vertical scale = horizontal scale
Pile
Pile
LEGEND
Fill
Alluvium sand
Surficial sand and terrace sediments
Pleistocene till units
Pleistocene sand and gravel (Empress Group)
Cretaceous bedrock (Lea Park Formation)

Figure 19. *Cross-section of the landslide at the Deer Creek bridge crossing of the North Saskatchewan River, Saskatchewan. The bridge is being deformed by active slip on two sliding surfaces within a complex landslide in Cretaceous shale (*modified from *Kelly et al., 1995, Fig. 9).*

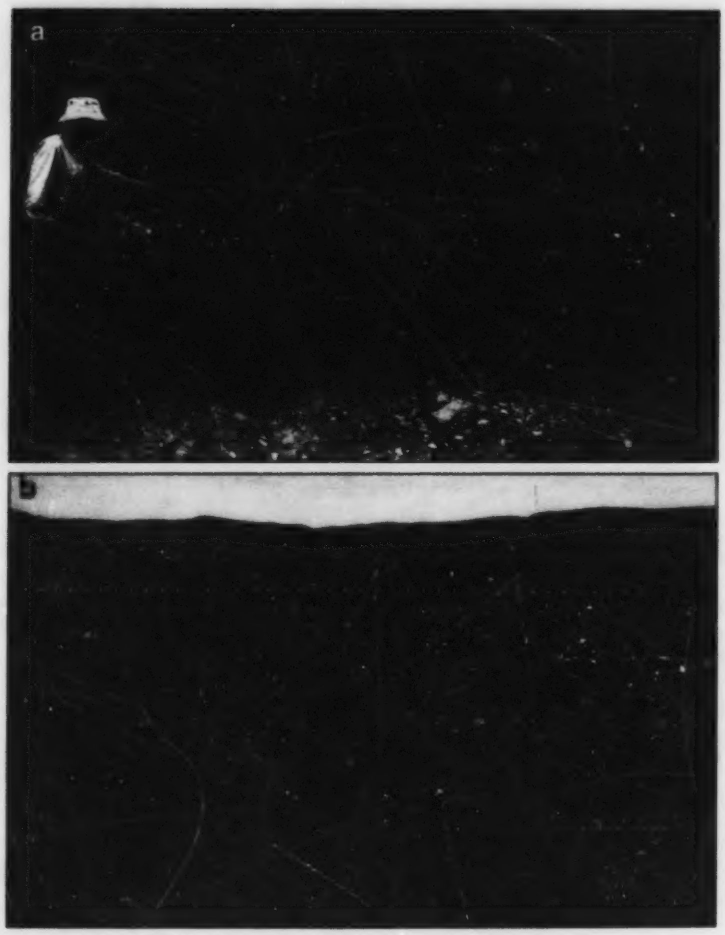


Figure 20.

a) *Varves in glaciolacustrine sediments, Thompson River valley near Ashcroft, southwestern British Columbia. Photograph by S.G. Evans. GSC 2000-183* ***b)*** *At about 01:00 h on 22 September 1897, residents of Ashcroft were awakened by loud, thunder-like rumblings produced by a landslide on the west bank of the Thompson River, just south of Ashcroft. Contemporary press reports indicate that the landslide was slow-moving and constricted the channel of the Thompson River without blocking it. The landslide appears to have been dominantly translational with the basal shear surface located in a glaciolacustrine unit. Photograph by S.G. Evans. GSC 2000-184* historical period. In 1997, also in the Fort St. John area, reactivation of a landslide in Cretaceous shale caused the rupture of a natural gas pipeline on the north side of the Peace River valley, 10.5 km upstream from the 1957 failure. The pipeline was relocated across the landslide and placed on wooden skids to accommodate possible future movement of the slope. The west abutment of the Little Smoky River bridge (Fig. 2b, 18), south of Peace River, Alberta, was founded on a slow-moving landslide. The structure of the bridge was modified by placing rollers under the superstructure to allow the abutment to move slowly towards the river without damaging the bridge.

On the eastern margins of the Cordillera, remarkable compound slides occur in Lower Cretaceous rocks in the Scatter River valley, Liard Plateau, northern British Columbia (Fig. 2b). Both sides of the Scatter River Canyon, which is over 400 m deep, are rimmed by a continuous zone of landsliding covering over 35 km². Resistant sandstone up to 80 m thick makes up the rimrock of the landslide complex and is underlain by weaker shale that dips about 4° towards the centre of the valley. Landslide blocks exhibit backward tilting, subsidence, and translation. The largest of the blocks measures 1.2 km by 0.3 km in plan and appears to have subsided about 100 m while moving 100 to 200 m horizontally and tilting slightly backward.

Slides are also the characteristic style of landslides in glaciolacustrine sediments deposited in glacial lakes that existed in many parts of Canada during the Pleistocene (Fig. 2a). Glaciolacustrine sequences are characterized by interbedded silt and low-strength clay sediments (Fig. 20a). They are commonly interlayered with sand partings in which excess pore pressures may develop. (The shear strength of geological materials is reduced by the pressure of water or gas that may be present in pores or voids within the material. High pore pressures commonly develop in sand layers within a finer grained sequence thus reducing the shear strength of the sandy materials.) Sensitive materials have been encountered in some glaciolacustrine sequences. (Sensitivity describes the relationship between in situ (or undisturbed) and remoulded (or disturbed), undrained shear strength. It expresses the ratio of undisturbed to remoulded, undrained shear strength as measured by apparatus in the field or the laboratory. Material is sensitive if values of this ratio exceed 1. Although its undisturbed, undrained shear strength may be substantial, the shear strength of disturbed sensitive material may very low. The loss of undrained shear strength during disturbance associated with failure, for example, may result in a material flowing like a liquid.) Failure geometries range from simple rotational slides to compound movements in which the sliding surface consists of multiple planar elements determined by the attitude of features such as fissures, laminations, and bedding (Fig. 20b). Many urban centres in the Prairie Provinces and British Columbia are built on glaciolacustrine sediments and landslides in these materials have affected sites along river banks and lake shores. Winnipeg, for example, is built on the former floor of Glacial Lake Agassiz and numerous slope stability problems have been encountered along the Red and Assiniboine rivers.

A vast array of techniques have been used to reduce slow slide movement to tolerable levels. These include the modification of slope geometry (most often through the construction of a stabilization berm at the toe of the slope) and internal drainage to reduce water pressures within the moving slope.

The gradual breakdown of slump blocks in slow slides may lead to the formation of slow earthflows downslope. In other cases, the catastrophic disintegration of sliding material may form rapid earthflows, also known as 'flowslides'.

SLOW EARTHFLOWS IN ROCKS AND QUATERNARY SEDIMENTS

Earthflows in rock generally develop as a result of the breakdown of slump blocks in the source material and commonly develop in materials, such as tuffaceous shale or serpentinite, that produce debris rich in smectite, an extremely weak clay.

Large-scale earthflows originating in bedrock materials have been reported from a number of locations in the Cordillera. For example, a number of large earthflows with volumes ranging from 10^6 m³ to 9.6×10^7 m³ occur along the western margins of the Interior Plateau of British Columbia, with the best known being the Drynoch Slide (Fig. 2b, 21).

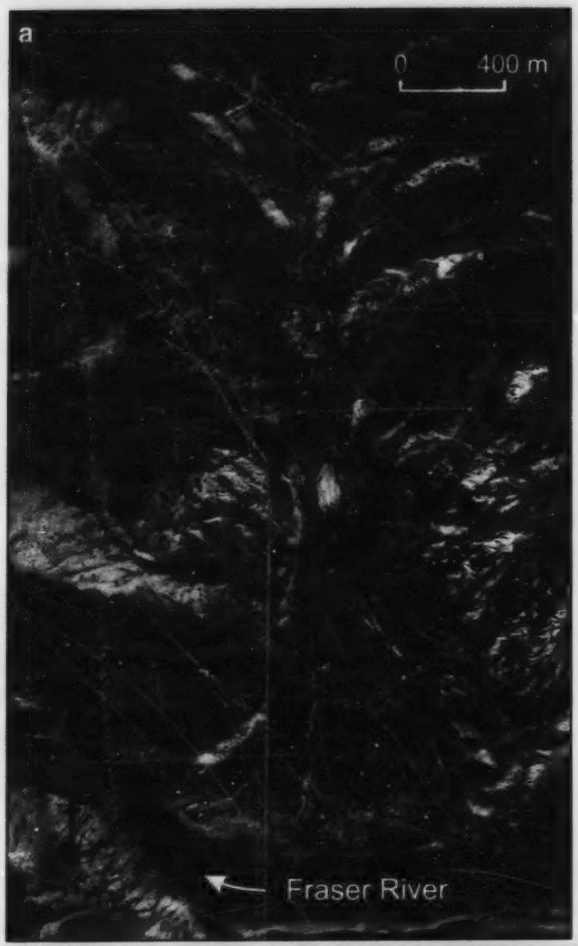
Figure 21. *Oblique aerial view of the Drynoch earthflow, Thompson River valley, British Columbia. The Trans-Canada Highway and the Canadian Pacific Railway track cross the toe of the landslide. Photograph by S.G. Evans. GSC 2000-088*

Earthflows along the Fraser River (Fig. 2b, 22a) have developed in Eocene volcaniclastic rocks. The degradation of these source materials has produced smectite-rich debris. Earthflows commonly show several phases of movement and the shear boundaries of these phases can often be clearly seen (Fig. 22a).

These landslides have been termed 'earthflows' despite the fact that movement involves mainly sliding on well defined shear planes. Detailed investigations at the Pavilion landslide, British Columbia (Fig. 2b, 22b), for example, show that although some internal deformation of the debris resulted from flow, movement is mainly by sliding on basal and lateral shears.

Earthflows in the Interior Plateau of British Columbia are slow-moving and show no evidence of rapid displacement. At the Pavilion earthflow, measured rates of movement are in the order of 45 cm/a; at the Drynoch Slide, estimated rates are in the order of 3.2 m/a for the period 1951 to 1972 (Fig. 1).

Earthflows commonly develop in the distal zones of landslides in glaciolacustrine sediments. At Big Slide, near Quesnel, British Columbia (Fig. 2b), retrogressive slide blocks have totally disintegrated to form an earthflow downslope (Fig. 23). Measurement of earthflow movement shows marginal sliding along well marked lateral shears, and internal deformation indicative of flow. Very high average annual rates of movement of up to 271 m/a (Fig. 1) have been measured in the Big Slide earthflow. These movements may have occurred in rapid surges of short duration.



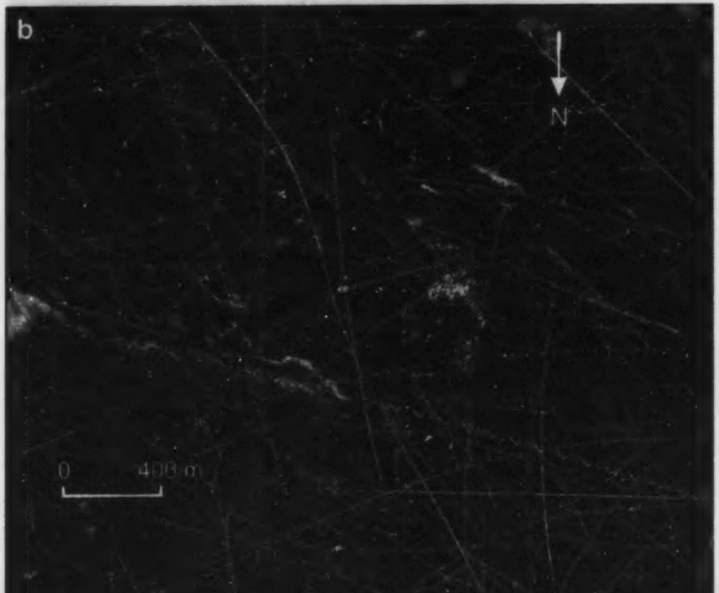
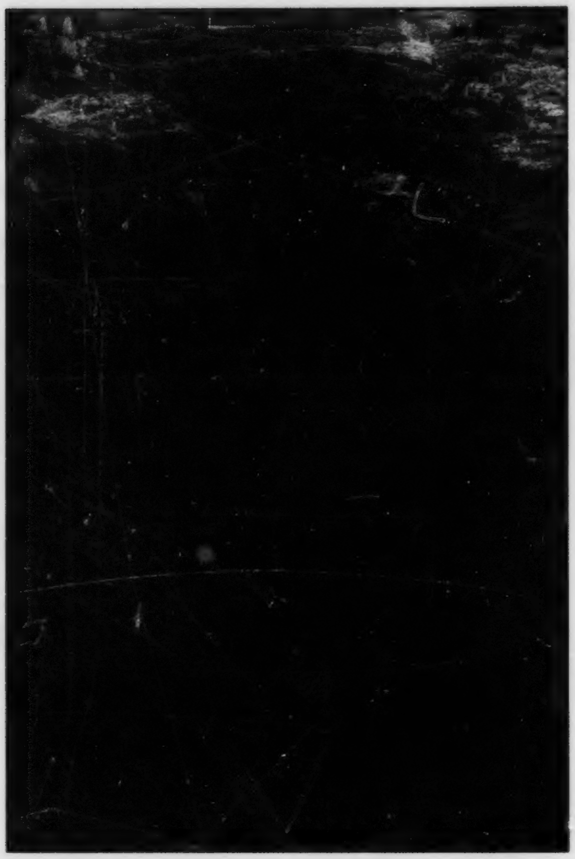


Figure 22.

***a)** Vertical aerial photograph of the Big Bar earthflow in Eocene volcanic and volcaniclastic rocks, Fraser River valley, Interior Plateau, British Columbia. Note the distinct lateral levees that define shear boundaries of flow. Province of British Columbia BC7705; 150 **b)** Vertical aerial photograph of the Pavilion earthflow, British Columbia. Note the source area, the lateral levees near the bottleneck, and the hummocky deposit on the valley floor. The earthflow is derived from Cretaceous continental sediments in the source area mixed with local till and colluvium. Province of British Columbia BC7788; 243*



RAPID SLIDES AND EARTHFLOWS (FLOWSLIDES) IN QUATERNARY SEDIMENTS

Glaciomarine sediments

Rapid landslides are especially common in the sensitive Pleistocene glaciomarine sediments of both Eastern and Western Canada (Fig. 2a). A number of significant landslides (Fig. 24) have occurred in the glaciomarine sediments (commonly called 'leda clay') of the St. Lawrence Lowlands of Eastern Canada during the historical period, with the earliest description dating from 1771. In addition, numerous scars of old landslides are found along river courses (Fig. 25) and terrace edges in the region. Glaciomarine sediments are derived from glacial rock flour and consist predominantly of clay-sized primary minerals (e.g. quartz, feldspar), although significant amounts of clay minerals (e.g. chlorite) may also be present. The particles form an open-structured fabric (sediments, e.g. glaciomarine, glaciolacustrine, and loess deposits, that have been deposited relatively rapidly, and have not been subject to intense overburden pressures, commonly exhibit an open-structured fabric in which the

Figure 23. *Aerial view of an earthflow in the South Bowl at Big Slide, near Quesnel, British Columbia. Note the well defined lateral shears. The earthflow originates in slump blocks in glaciolacustrine sediments at the head of the landslide complex. Surface movements as high as 271 m/a have been measured on the earthflow. Photograph by S.G. Evans. GSC 2000-089*

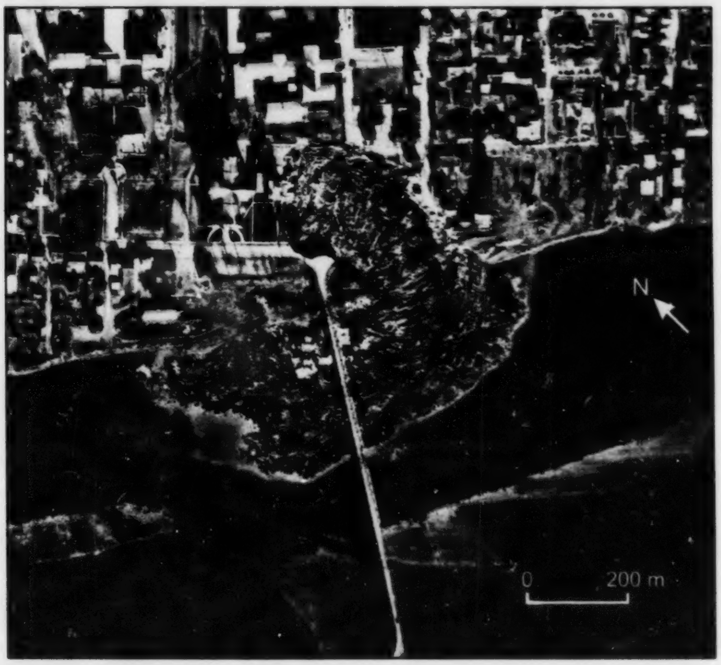


Figure 24.

Aerial photograph of a rapid earthflow in leda clay at Nicolet, Quebec, that took place on 12 November 1955. Three people were killed in the landslide that occurred quite rapidly over a period of two to three minutes and partly destroyed a church complex. The 147-year-old Cathedral of Saint-Jean-Baptiste, which stood on the edge of the crater had to be demolished following the landslide. Total damage was estimated to be in the order of $10 million (1955 dollars). Province of Quebec 377-24

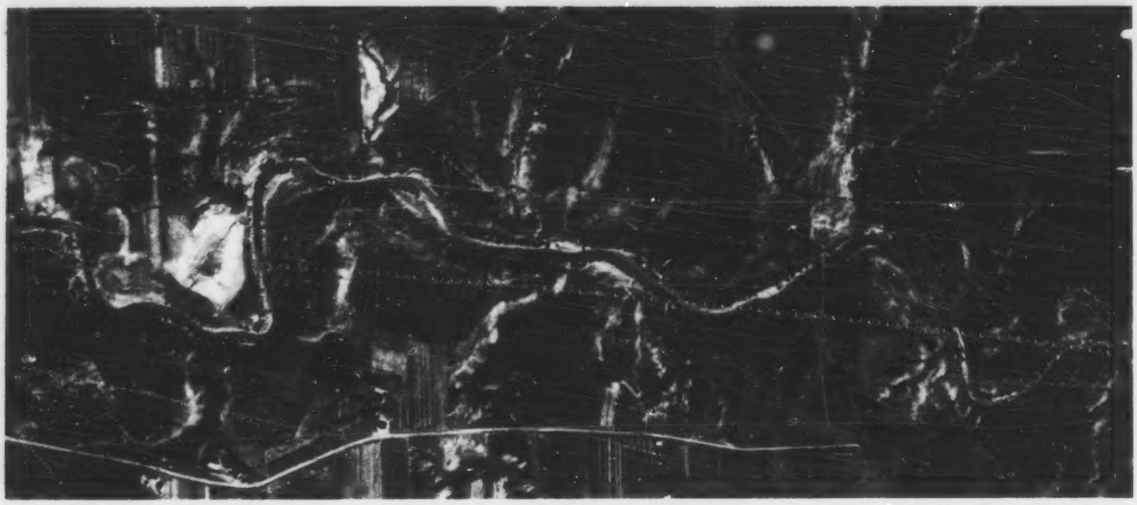
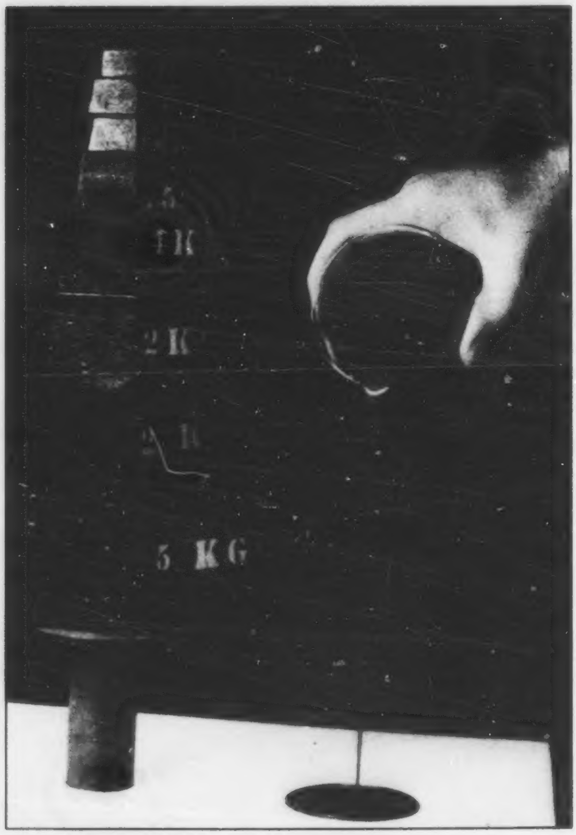


***Figure 25.** Earthflow scars in leda clay along the Champlain River, Quebec. The arrow shows the location of the 1895 Saint-Luc-de-Vincennes landslide in which five people lost their lives. NAPL A14600-29*



constituent soil particles have significant void spaces between them; the structure is maintained by contacts between the particles, which may be chemically cemented) commonly sustained by cementation. Materials with an open-structured fabric can become unstable when physically disturbed (e.g. by vibration or initial landslide movement; landslides in sensitive glaciomarine sediments have been triggered by blasting), resulting in significant loss of strength. Such materials have high sensitivities.

The behaviour of leda clay landslides reflects the complexity of the glaciomarine source materials, which were initially deposited in salt or brackish water, and the extent to which their chemistry has been modified by postdepositional processes. Initial failure may be followed by a process of retrogression that is dependent on both the initial strength of the material and its ability to disintegrate and flow after failure, which is in turn a function of its loss of strength or sensitivity (Fig. 26). The flowing of part of the failed material out of the landslide crater is a critical component of the retrogressive process, since the removal of material from the base of the landslide scarp destabilizes the intact slope, causing successive retrogressive failures. The development of rapid retrogressive earthflows has been found to be associated with the

←

***Figure 26.** The sensitivity of leda clay illustrated in samples from Ottawa. The sample on the left is intact. This undisturbed sample is strong enough to withstand a vertical load of 11 kg, a vertical stress of more than 100 kPa. A sample of the same material is being poured at right after having been remoulded at the same moisture content and behaves like a fluid. Photograph courtesy of the National Research Council of Canada*

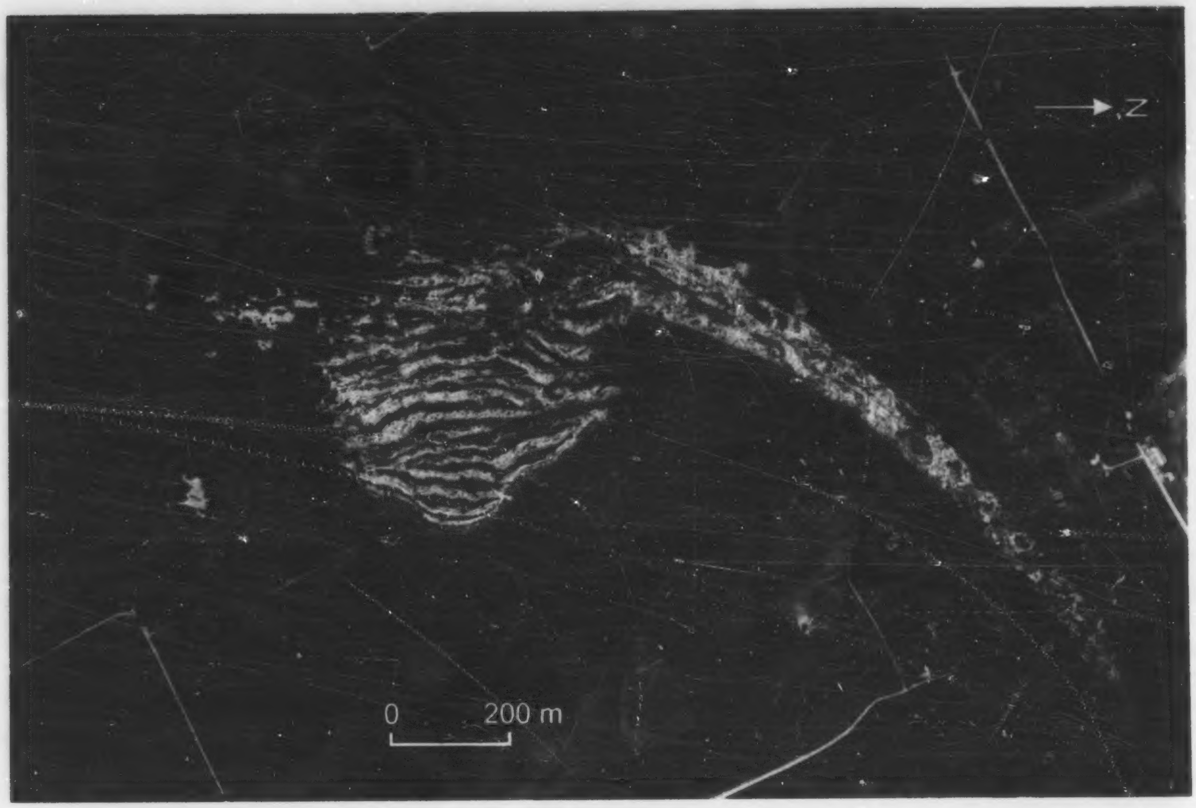


Figure 27. Aerial photograph of the 1971 South Nation River landslide, eastern Ontario. The earthflow occurred in leda clay on 16 May 1971. Debris flowed into the channel of the South Nation River and blocked its flow, causing flooding upstream that is still evident in this photograph taken on 1 September 1971. The volume of the landslide was approximately 7 x 10^6 m^3. NAPL A30362-152

presence of zones of sensitive clay at the base of the initial slopes. In turn, high sensitivity is associated with the reduction of the original salt content by groundwater leaching.

Initial failure mechanisms in glaciomarine sediments are complex and may involve spreading (i.e. the extrusion of disaggregated material under natural cap loading) as well as rapid sliding. In some cases, initial failures develop into rapid, highly retrogressive earthflows. In this type of landslide, rapid retrogression occurs and may extend hundreds of metres from the initial failure. At the same time, a significant part of the clay and silt in the failed mass is fluidized and flows rapidly in the opposite direction into an adjacent river channel. The debris can form a landslide dam that blocks river flow causing flooding upstream (*see* Brooks et al., 2001; Fig. 27). Most of the material that is not fluidized remains in the landslide crater, forming characteristic linear ridges and pyramidal blocks of intact clay (Fig. 27, 28). Some intact blocks of sediment float out of the crater in the flowing, liquefied clay.

In spite of the fact that landslides in leda clay involve limited original slope heights (usually less than 30 m), both the retrogression and flowage phases of the landslides can be extremely rapid and thus very destructive. In some cases, survivors of retrogression have had to run to avoid being swallowed up by the debris, which indicates retrogression speeds of approximately 5 m/s or greater. For example, at the 1894 Saint-Alban, Quebec, landslide (Fig. 2b), survivors report that in some places the failing earth was moving faster than they could run in the opposite direction (Anonymous, 1894). A similar experience occurred at the 1971 Saint-Jean-Vianney, Quebec, landslide (Fig. 2b, 29), in which a man running to escape the rapidly retrogressing failure reported having to run on what seemed like moving stairs, indicating that the final stages of the retrogression matched his running speed. It is noted that all 31 deaths at Saint-Jean-Vianney occurred when houses were engulfed in the retrogression phase of the landslide (Fig. 29).

The movement of the disintegrating landslide debris may also be extremely rapid and a number of deaths have occurred when victims were overwhelmed by fast-flowing leda clay debris. Eyewitnesses estimated that the clay blocks carried along on the surface of the flowing clay in the 1898 Saint-Thuribe landslide, Quebec (Fig. 2b) were moving as "fast as a horse can gallop", i.e. approximately 10 to 13 m/s.

Figure 28.

View of leda clay debris in a crater formed by the 1993 Lemieux landslide, South Nation River, Ontario. The earthflow flowed towards the left of the photograph. Note the intact blocks and ridges as well as remnants of the former grassy surface. The people provide scale. Photograph by S.G. Evans. GSC 2000-090

Where the distal flow is channelized, velocities may be estimated from the superelevation of the debris surface as it passed through bends. In the Saint-Jean-Vianney landslide, the distal flow travelled through the narrow des Vases River down to the Saguenay River at approximately 7 m/s (25 km/h; Fig. 1), destroying a bridge in its path and carrying the bridge's concrete piers into the Saguenay River (Fig. 29).

Mitigation of leda clay landslide hazard in Eastern Canada has consisted of preventing initial failure by constructing protective berms at the toes of slopes and by establishing setbacks, no-go hazard zones along the tops of slopes defined on the basis of the calculated maximum distance of retrogression should failures occur. Development within these hazard zones is restricted and, in some instances, as in the case of Lemieux, Ontario (Fig. 2b), existing buildings have been moved and the properties expropriated after the zone has been defined.

Landslides in leda clay may involve substantial volumes of material. Approximately 17 historical landslides involving over 10^6 m^3 of material have occurred between 1840 and 1998. The largest is the remarkable 1894 Saint-Alban landslide in Quebec, which involved an estimated 185 x 10^6 m^3 of material; the landslide scar covers 462 ha and has an average depth of 40 m. Four people were killed in the landslide as farmhouses were engulfed in the debris. Landslide debris blocked the Sainte-Anne-de-la-Pérade River. When the debris dam broke, an outburst flood swept down the river, destroying several bridges. Although the movement took place in multiple phases over a three-hour period, the landslide is arguably the largest landslide known to have occurred in Canada between 1840 and 1998.

In 1993, a large landslide occurred in leda clay at Lemieux, Ontario (Fig. 2b), and temporarily blocked the South Nation River (Fig. 30a). The hazard had been previously recognized and the village of Lemieux had been moved prior to the landslide. In April 1996, the largest leda clay landslide of the twentieth century occurred at Saint-Boniface-de-Shawinigan, Quebec (estimated volume 7–8 x 10^6 m^3; Fig. 2b, 30b). Landslides in leda clay occur most commonly in the spring during melting of the winter snowpack.

Rapid slides and earthflows have also occurred in the less well known sensitive glaciomarine sediments of coastal British Columbia. In January 1880, a major landslide (estimated volume ~10^6 m^3) occurred at Haney, on the north bank of the Fraser River, in the Lower Mainland of southwestern British Columbia (Fig. 2b). The slope involved in the failure is about 30 m high and retrogression was in the order of 250 m. Debris partly blocked the Fraser River, which at this point is about 475 m wide, obstructing river traffic and forcing Aboriginals to portage their canoes around the turbulent channel, which "ran like a mill race". Approximately 8 ha of land were displaced very rapidly downward and outward and the debris generated a deadly displacement wave in the Fraser River.

More recently, about Christmas 1993, a large earthflow covering 23 ha and involving 1.4 x 10^6 m^3 of material occurred in sensitive glaciomarine sediments near Terrace, northwestern British Columbia (Fig. 2b). Spoil from the crater filled the Mink Creek valley for about 1 km, damming the river and forming a lake that extended 1200 m upstream. The average depth of the basal failure surface was 6 m and the slide retrogressed a maximum of 400 m. Debris in the crater contains numerous ridges and remarkable pyramidal blocks, similar to debris from sensitive clay landslides in Eastern Canada. The Mink Creek earthflow is one of numerous flowslides to have occurred during the Holocene in the glaciomarine sediments of the Terrace–Kitimat area.

Glaciolacustrine sediments

Rapid slides and, less commonly, rapid earthflows may also occur in glaciolacustrine sediments. The mechanisms for the development of such landslides in these sediments are not yet fully understood. A striking example of a rapid earthflow (flowslide) in glaciolacustrine sediments is the 1973 Attachie landslide in the Peace River valley, 50 km west of Fort St. John, British Columbia (Fig. 2b). The total volume of material involved in the slide was about 12.4 x 10^6 m^3 (Fig. 31). Sliding was seated in silty clay glaciolacustrine material underlying a cap of glacial till and took place on a valleyside that had undergone considerable prefailure displacement. About 6 x 10^6 m^3 of material was left on the failure surface as slide blocks, and the remainder (6.4 x 10^6 m^3) formed a high-speed flow that travelled almost 1 km across the flat-bottomed Peace River valley (Fig. 31). The distal flow had an estimated emplacement velocity of 20 m/s (72 km/h). Similar high-velocity flowslides occurred in silty glaciolacustrine sediments in the Thompson River valley of southwestern British Columbia in 1880 at Ashcroft (estimated volume 15 x 10^6 m^3; Fig. 2b) and in 1908 at Spences Bridge (described below; Fig. 2b).

In Eastern Canada, similar landslides have occurred in glaciolacustrine sediments that have been shown to be sensitive (e.g. 1990 Nipigon River landslide, Ontario; Fig. 2b). In the case of the 1946 Beattie mine disaster at Duparquet, in western Quebec (Fig. 2b), a flow generated in sensitive glaciolacustrine materials around the rim of the open pit, or 'Glory Hole', became fluid enough to penetrate underground workings of the mine where four miners were killed by the flowing debris.

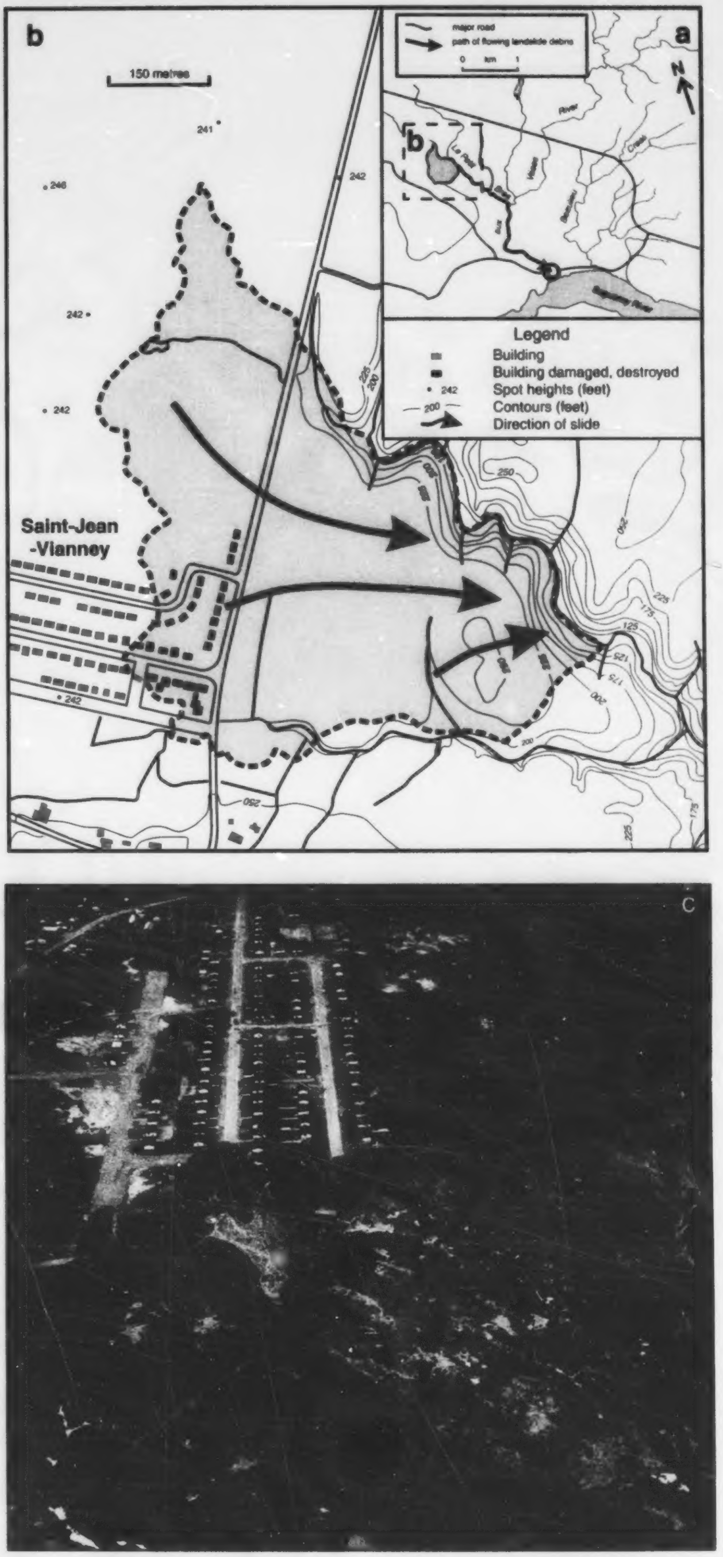

b
150 metres
241
246
242
242
242
242
Saint-Jean
-Vianney
225
200
250
250
225
175
125
200
a
major road
path of flowing landslide debris
0 km 1
N
b
River
Creek
Saguenay River
Legend
Building
Building damaged, destroyed
242 Spot heights (feet)
200 Contours (feet)
Direction of slide
c

Figure 29.

Saint-Jean-Vianney landslide, Quebec, 4 May 1971. ***a)*** *Location map of the landslide, a rapid earthflow in leda clay (from NTS 22D/06). Note the location of the landslide scar. The circle marks the location of a bridge destroyed by rapidly moving, liquefied debris as it swept down the aux Vases River to the Saguenay River.* ***b)*** *Large-scale map of the landslide scar showing the limit of retrogression and houses that were engulfed in the debris during the rapid retrogressive phase.* ***c)*** *Oblique aerial view to the west of the landslide headscarp. Approximately 40 houses were engulfed and 31 people died in the leda clay earthflow, Canada's worst postwar landslide disaster. Some displaced houses are seen in the debris below the scarp. The flow moved towards the bottom right. (Photograph taken by Canadian Forces reconnaissance aircraft on 7 May 1971, courtesy of R. Bergeron, ministère de l'Énergie et des Ressources, Quebec, and Canadian Forces, Bagotville).*

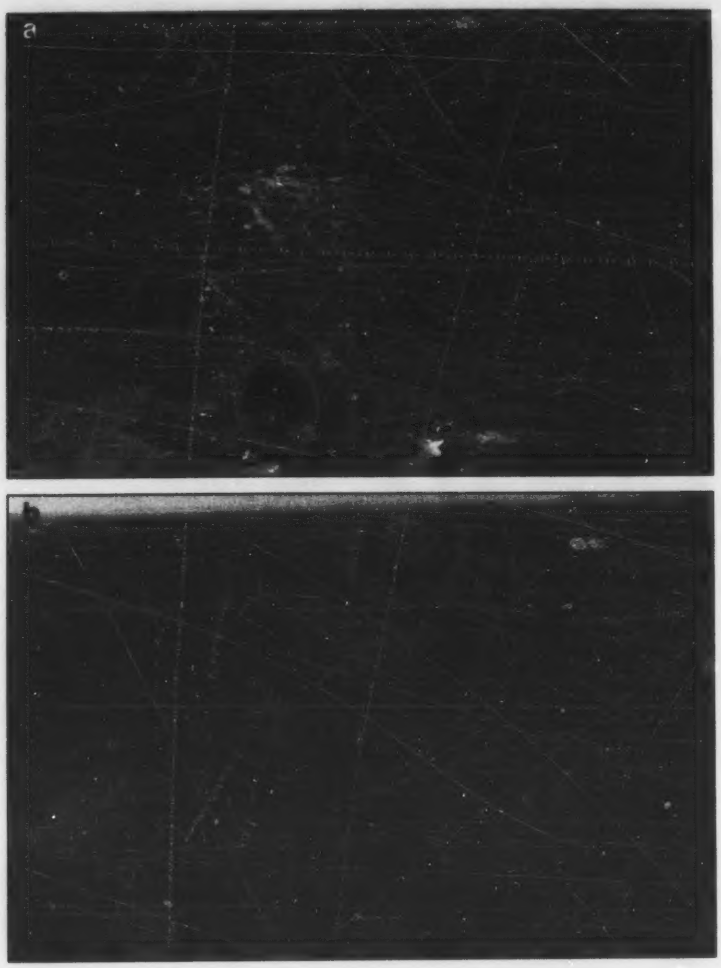


Figure 30.

a) Oblique aerial photograph of the 1993 Lemieux landslide, South Nation River, eastern Ontario. The rapid earthflow in leda clay involved 2.8 x 10^6 m^3 of material and retrogressed 680 m from the edge of the river. Debris from the landslide flowed into the channel of the South Nation River, forming a temporary dam. Photograph by S.G. Evans. GSC 2000-091
b) Oblique aerial photograph of the landslide that occurred in sensitive leda clay at Saint-Boniface-de-Shawinigan, Quebec, in 1996. Note the ponded waters of the Machiche River and the retrogressive slide ridges. The landslide covers about 30 ha and involves a volume of about 7 x 10^6 m^3 to 8 x 10^6 m^3 of material. It retrogressed along a 1.1 km front, but did not form a significant distal flow. Photograph by S.G. Evans. GSC 2000-092

Figure 31.

Oblique aerial view of the Attachie landslide that occurred in 1973 on the southeast wall of the Peace River trench, 50 km west of Fort St. John, northeastern British Columbia. View is downstream towards the east. The landslide involved a slope consisting of glaciolacustrine sediments capped by glacial till. The total volume of the displaced mass is estimated to be 12.4 x 10^6 m^3 of which 6.4 x 10^6 m^3 formed a rapid flowslide that swept across the floor of the Peace River valley. The debris blocked the flow of the Peace River for about 10 hours. Photograph courtesy of Aurora Color Studio, Fort St. John, British Columbia.

LANDSLIDES IN PERMAFROST TERRAIN

Over 50% of Canada's land mass is underlain by permafrost (*see* Smith et al., 2001). In these regions, the presence of ground ice is a major factor affecting slope stability either because of the presence of a significant thickness of frozen materials in a slope, which produces a material with distinctive geotechnical properties, or because of the melting of previously frozen ground. The impact of permafrost on slope stability is most marked in slopes consisting of Quaternary materials (glacial till, colluvium, glaciolacustrine and glaciomarine materials). As discussed below, the freezing and melting of materials in slopes are strongly controlled by climatic factors and the extent to which natural or anthropogenic disturbance affects the heat flux in the upper part of the geotechnical soil profile.

Frozen slopes

In permafrost areas, an upper unit of frozen sediment may create a rigid material that forms a resistant cap in some slopes. The frozen cap has greater strength than the unfrozen material beneath it and thus forms the steeper upper part of a partially frozen slope. Landslides occur in such slopes under conditions of active toe erosion. In a coastal environment, erosion at sea level may undercut frozen slopes. Along eroding river banks, resistant frozen caps may create loading conditions favourable for the failure (Fig. 32) of underlying, unfrozen sediments and the subsequent development of retrogressive slides.

Melting slopes

Earthflows are common in ice-rich silt and clay slopes in permafrost areas (Fig. 33), where failure commonly occurs in the active layer during summer thaw. Two types of earthflow have been distinguished, skin flows (or active-layer detachments), which involve the detachment of a thin veneer of vegetation and soil and subsequent movement over the inclined surface of the permafrost table, and bimodal flows (or retrogressive thaw flows), which consist of a steep headscarp and a low-angled tongue. Skin flows are generally shallow. Bimodal flows are deeper seated and exhibit a characteristic bowl shape defined by a marked scarp containing ice-rich

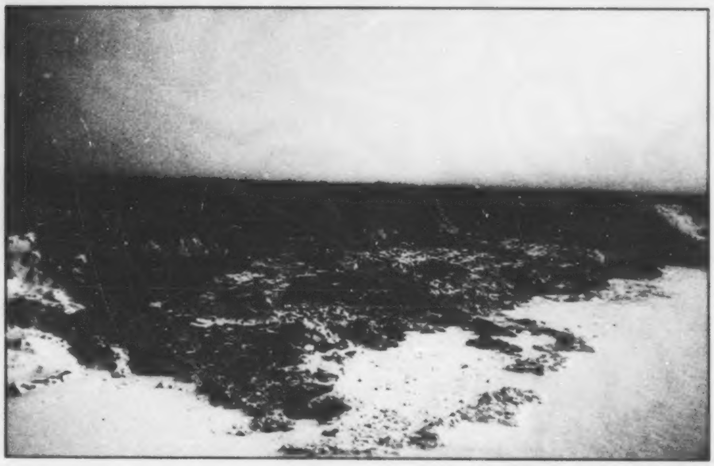


Figure 32.

Landslide that occurred in 1997 in frozen glaciolacustrine sediments in the bank of the Mackenzie River, south of Old Fort Point, Northwest Territories. Photograph by M.M. Burgess GSC 2000-093

Figure 33.

Earthflow in permafrost terrain, Dekale Creek, Mackenzie Mountains, Northwest Territories. Photograph by S.G. Evans. GSC 2000-094 sediments. Under certain conditions, both types of landslide may evolve into more extensive complexes as thawing of the slide mass and/or adjacent ice-rich sediments occurs; the Surprise Rapids Landslide complex in the Yukon Territory, for example, covers 1.7 km² (Fig. 2b). These types of landslide are important processes in permafrost degradation following thermal or physical disturbance as, for example, in the case of forest fires (Fig. 34).

Landslides involving melting permafrost may take place on low-angle slopes as a result of the generation of excess pore pressures during thaw consolidation.

In high alpine areas south of the permafrost limit and in mountain areas within the permafrost limit, including the Arctic Islands (e.g. Sawtooth Range, Ellesmere Island; Fig. 2b), the melting of segregated ice, including buried glacier ice in end-moraine complexes, may mobilize debris flows. One such debris flow was observed at Steele Creek (Fig. 2b), in the St. Elias Mountains, Yukon Territory, in 1967. It exhibited pulses of movement and transported boulders up to 4 m in diameter.

Figure 34. *Landslides on the bank of the Mackenzie River, near the mouth of Thunder River, Northwest Territories, caused by the degradation of permafrost due to a forest fire in 1986. Retrogressive thaw flows in the upper part of the slope generate saturated debris that flows down to the river. Photograph by J.M. Aylsworth, taken in 1994. GSC 2000-095*

LANDSLIDES IN WASTE DUMPS, EMBANKMENTS, AND EXCAVATED SLOPES (GEOTECHNICAL FAILURES)

Significant landslides can result from the failure of man-made road and rail embankments, excavated rock slopes, and mine waste dumps. These types of landslides have been termed 'geotechnical failures'. Although they are not natural landslide events *sensu stricto*, they do involve reworked natural materials and are commonly caused by extreme hydrometeorological conditions (e.g. heavy rainfall, rapid melting of snowpack) or earthquake shaking.

Man-made dumps and embankments

Some of the largest landslides in Canada in recent years have occurred on the slopes of gigantic waste dumps at open-pit coal mines in the Rocky Mountains of southeastern British Columbia where 10 mines currently generate 200 x 10^6 to 250 x 10^6 m³ of rock waste each year. The waste is disposed of in huge dumps constructed by end tipping from dump trucks (Fig. 35). Over 50 high-velocity, long-runout flowslides resulting from failures in these dumps have been documented since 1972; at least nine flowslides involving over 10^6 m³ of material occurred in the period 1971 to 1993, culminating in the gigantic 8 x 10^6 m³ failure of Fording Coal's South Spoil in the Fording River valley in 1993 (Fig. 2b). Peak velocities in this type of flowslide may reach 40 m/s (144 km/h). Failure in the dumps is closely monitored and complete slope failure can be predicted using deformation–time plots. Coal-mine waste-dump failures have claimed nine lives in British Columbia in the period 1968 to 1996.

Figure 35.

Dump truck driving away from the edge of a waste dump after end-dumping rock waste at open-pit coal mine in the East Kootenay Coalfields, southern Rocky Mountains, southeastern British Columbia. Individual dumps can reach 400 m in height and fail forming large-scale, rapid flowslides that may reach velocities of 40 m/s and travel more than 2 km from the base of the dump slope. Note tension cracking near the dump truck, which indicate movement in the dump; these cracks at the dump crest are closely monitored and accelerating displacements forewarn of failure. Photograph by S.G. Evans. GSC 2000-096

Rapid landslides also occur in road and railway embankments, often with disastrous results (Fig. 36). Failure may take place within the embankment fill itself, or involve basal failure in the embankment foundations (Fig. 37).

Excavated slopes

Major landslides have occurred in excavated slopes associated with large-scale construction projects and open-pit mining in Canada. At the Revelstoke dam (Fig. 2b), on the Columbia River, for example, several complex rock-slope movements were encountered during excavations at the site between 1977 and 1980. At the diversion tunnel inlet, a catastrophic failure (estimated volume 12 000 m³) occurred in heavily fractured metamorphic rocks associated with the Columbia River Fault Zone. Sliding took place along an undulating foliation shear zone containing brecciated, clay-rich gouge. In addition, during excavations for the dam and powerhouse, movements were detected in both the east and west walls of the excavation and were stabilized using a variety of stabilization techniques.

Figure 36. *Aerial oblique view of a railway embankment failure that occurred at Conrad Siding, near Lytton, in the Fraser Canyon of southern British Columbia, in March 1997. A freight train plunged into the pit produced by the failure, resulting in the death of two crewmen. The failure also undermined the Trans-Canada Highway, visible at the top of the photograph. The saturated fill material formed an earthflow that flowed down to the Fraser River in the foreground. Photograph by S.G. Evans. GSC 2000-097*

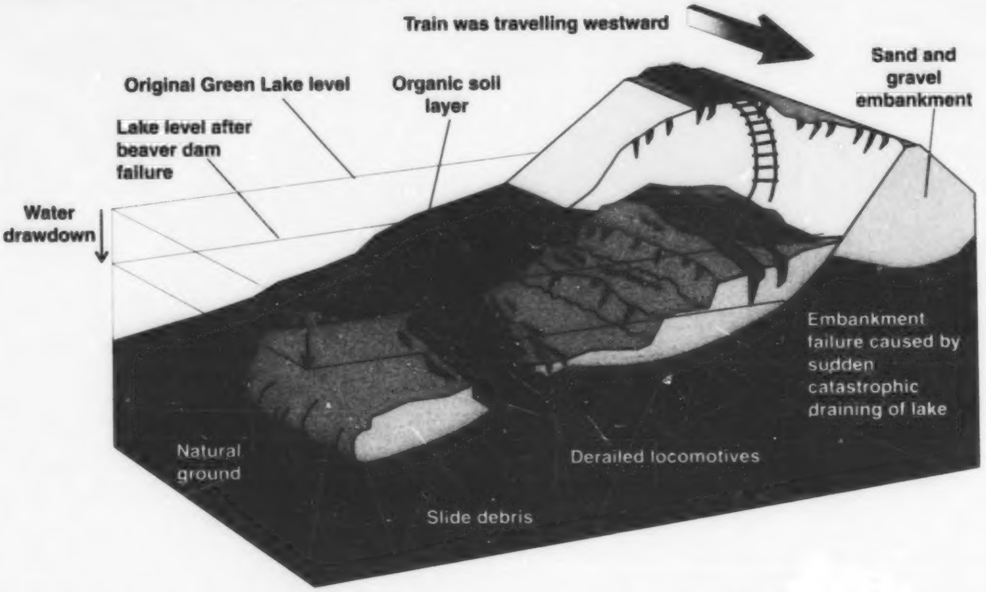

Train was travelling westward
Sand and gravel embankment
Original Green Lake level
Organic soil layer
Lake level after beaver dam failure
Water drawdown
Embankment failure caused by sudden catastrophic draining of lake
Natural ground
Derailed locomotives
Slide debris

Figure 37. *Artist's impression of an embankment failure that led to the derailment of a freight train at Nakina, Ontario, in July 1992. The embankment failure resulted from the rapid drawdown of a beaver-dammed lake when the beaver dam failed. Two railway employees lost their lives in the derailment* (modified from *CN North America, 1993*).

Rock-slope failures in open-pit mines have been widely documented in Canada and inevitably occur as part of the process of profitable mining. They range from complex failures involving sliding and toppling to relatively simple footwall failures in open-pit coal mines. During the 1970s, complex movements took place in the Jeffrey open pit mine at Asbestos, Quebec (Fig. 2b), and involved both bedrock and overburden slopes. In the January 1975 failure, a landslide (estimated volume 1.3 x 10^6 m^3) in Pleistocene sediments on the east rim of the mine retrogressed beyond the mine property and into an adjacent residential area, leading to the relocation of 172 families. Debris from the landslide partly filled the pit and caused losses in mine production estimated at \$4–5 million (1975 dollars). One of the largest catastrophic open-pit failures documented in Canada (estimated volume ~2 x 10^6 m^3) took place in 1990, in a 333 m high slope at the Brenda mine near Peachland, British Columbia (Fig. 2b, 38). Failure was sudden and the impact of the debris on the floor of the open pit generated a seismic shock of local magnitude 2.4 that was measured on seismographs up to 215 km away.

At the Syncrude mine (Fig. 2b) in the Athabasca Tar Sands, Alberta, open-pit mining by gigantic draglines necessitates the excavation of high (about 50 m), steep slopes directly adjacent to and below the machines (Fig. 39). Slope stability considerations are paramount in both the economies of mining and the security of the draglines. Instability results from rapid block slides along clay seams within the tar sands and more ductile failure due to gas exsolution in the bitumen-saturated sand. Syncrude has developed sophisticated monitoring technology to warn of excessive movement in the highwall that may signal potential danger to the dragline.

UNDERWATER LANDSLIDES

Underwater landslides occur on the steep slopes of marine and lacustrine deltas as well as on the slopes of Canada's continental shelf. High sedimentation rates in these environments maintain steep underwater slopes that are subject to mass failure; this is a hazard for port facilities, buildings sited along shorelines, and underwater electricity and communication cables.

Slides on submarine delta slopes are commonly triggered by excess pore pressures generated by the drawdown associated with low tides. At Woodfibre (Fig. 2b), on the western side of Howe Sound in southwestern British Columbia, for example, an underwater landslide on the delta front in 1955 caused \$0.5 million to \$0.75 million (1955 dollars) in damage to wharves and warehouses sited on the delta. The failure took place a few minutes after extreme low tide.

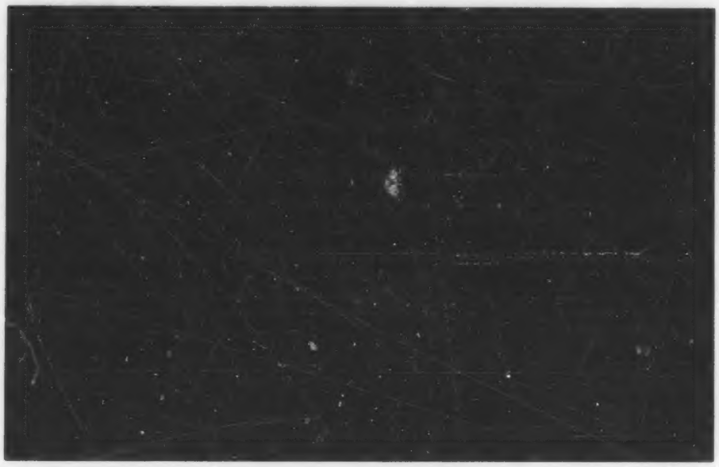


Figure 38.

The 1990 failure of the west wall of the Brenda mine, near Peachland, British Columbia. The double-benched wall configuration with catch berms at 100 foot (30.7 m) intervals gives scale. Water leaked into the pit from a broken water pipe after the failure. Relief from the top of the pit to the water surface is approximately 300 m. Photograph by S.G. Evans. GSC 2000-098

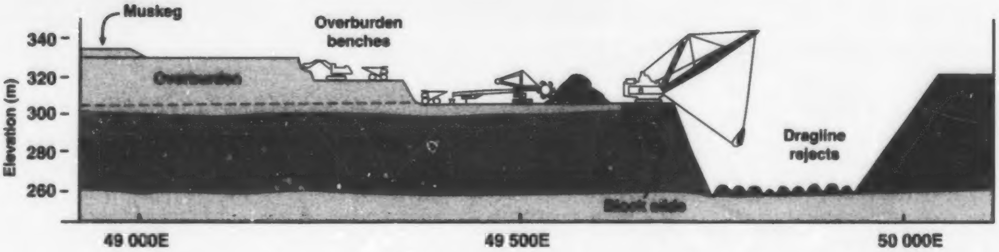


Figure 39. *Cross-section of mining operations at the Syncrude Pit, Athabasca Tar Sands, Alberta, showing the location of a typical blockslide in the McMurray Formation in relation to massive draglines that mine the bitumen-impregnated sands* (modified from *Morgenstern et al., 1988).*

Large-scale delta-front failures have been documented in a number of British Columbia fiords. In 1975, a series of massive submarine landslides occurred in the vicinity of the Kitimat River delta front at the head of Kitimat Arm, a fiord on the northern coast of British Columbia (Fig. 2b). Some question remains concerning the exact sequence of submarine landslide events on 27 April. It is known, however, that two major failures were involved, i.e. a landslide in sensitive marine clay on the west wall of the fiord and a large-scale failure of deltaic sediments from the foreslope of the Kitimat River delta (Fig. 40). The sequence of events was most likely initiated by the landslide on the fiord wall, which was triggered by construction activity on the shoreline. The failed mass of delta and sidewall sediment (estimated volume 10.5 x 10^6 m^3) slid down into the fiord and apparently caused the mobilization, by undrained loading, of a further 25 x 10^6 m^3 of fiord-bottom sediments, creating a massive, submarine, kilometre-wide debris flow of about 35 x 10^6 m^3 that travelled over 5 km from the delta front over slopes of less than 0.4° to a water depth of 210 m. Individual blocks (outrunner blocks) of considerable size (up to 35 000 m^3) slid along the seafloor up to 1 km beyond the distal limit of the main flow (Fig. 40). The landslides generated a series of displacement waves that caused substantial damage to docking facilities in the upper reaches of Kitimat Arm, as described below.

In southwestern British Columbia, recurrent submarine flowslides are a natural process in the progradation of the Fraser River delta (Fig. 2b), where the presence of interstitial methane gas in loose, newly deposited sediments is an important factor in triggering underwater failures. In 1985, a large-scale, submarine flowslide (estimated volume 10^6 m^3) retrogressed to within 100 m of the Sand Heads lighthouse, on the southern part of the delta.

In Eastern Canada, the massive 1929 Grand Banks failure (Fig. 2b) has been the subject of considerable investigation and much discussion. The events were associated with a M7.2 earthquake with an epicentre located at the extreme western margin of the Grand Banks, about 280 km south of Newfoundland. The location of the epicentre coincided with an area of massive submarine failure on the continental slope that resulted in a number of instantaneous breaks of telephone cables on the day of the earthquake. Recent work has established that the seismic event was generated by a true tectonic earthquake involving faulting in the crust and not by the massive landslide itself. The timing of cable breaks on 18 November 1929 and subsequent ocean-floor mapping define an area of simultaneous, earthquake-triggered slope failure over 20 000 km^2 on the upper part of the continental slope. The initial landslide involved Holocene and Pleistocene muds that,

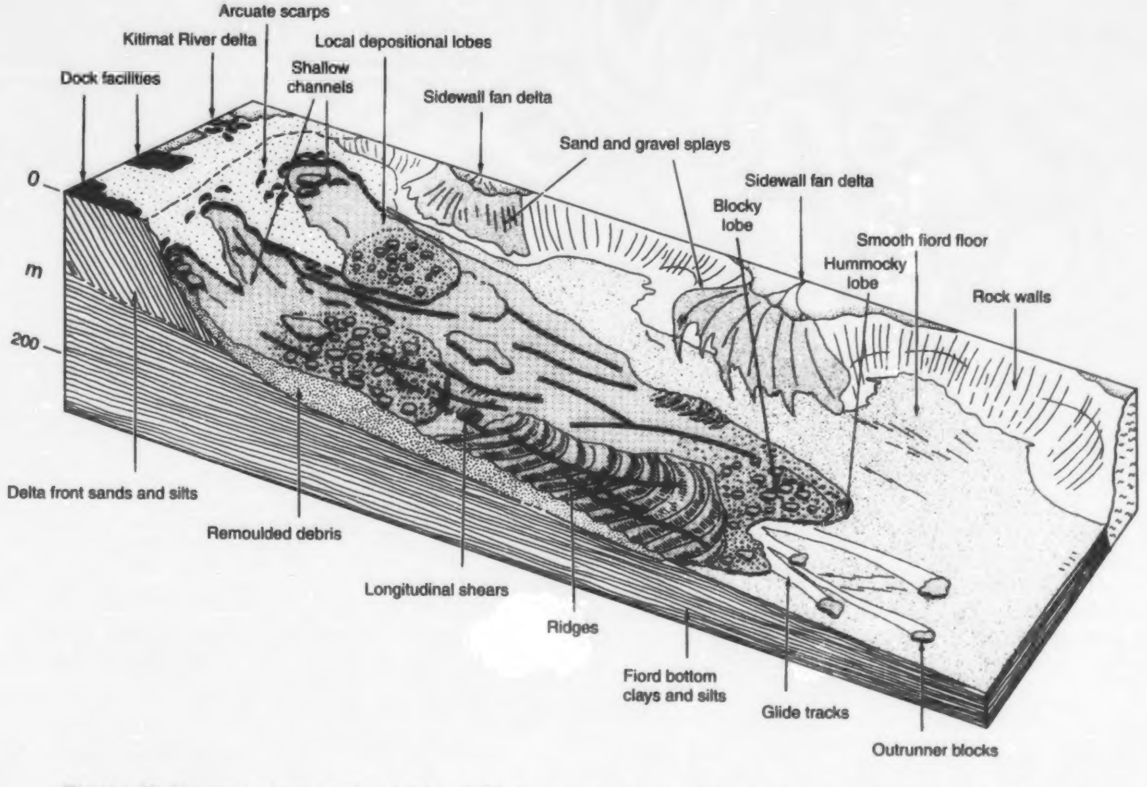

Arcuate scarps
Kitimat River delta
Local depositional lobes
Dock facilities
Shallow channels
Sidewall fan delta
Sand and gravel splays
Sidewall fan delta
0
Blocky lobe
Smooth fiord floor
m
Hummocky lobe
Rock walls
200
Delta front sands and silts
Remoulded debris
Longitudinal shears
Ridges
Fiord bottom clays and silts
Glide tracks
Outrunner blocks

Figure 40. *Kitimat submarine landslide, 1975. Schematic sketch of the seafloor morphology at the head of Kitimat Arm illustrating the geomorphology of failures on the Kitimat River delta and submarine debris flow* (modified from *Prior et al., 1984, Fig. 2).*

in moving down the continental slope, entrained coarser sediments. Ultimately, the landslide was transformed into a sandy turbidity current that travelled over 700 km along the floor of the Atlantic into the depths of the Abyssal Plain. On the basis of the timing of cable breaks on the continental slope, the submarine flow reached velocities of at least 19 m/s (67 km/h). It is estimated that about 10^{11} m^3 of material were displaced in the submarine landslide, a volume over 500 times greater than that of Canada's largest known historical subaerial landslide, the 1894 Saint-Alban event. The instantaneous failure of this massive volume of material generated a deadly tsunami in the Atlantic Ocean that swept ashore on the Burin Pensinsula (Fig. 2b) of southern Newfoundland (described below; *see also* Clague, 2001).

THE SECONDARY EFFECTS OF LANDSLIDES

The secondary effects of landslides (landslide-generated waves and landslide dams) can cause damage in areas beyond the limits of landslide debris and can be very destructive.

Landslide-generated waves

Where rapid landslides enter rivers, lakes, artificial reservoirs, or the sea, they may generate destructive water waves (also known as 'landslide tsunamis'), which have caused substantial damage in Canada in the historical period, as in the case of the 1905 landslide at Spences Bridge, British Columbia, and the 1908 landslide disaster at Notre-Dame-de-la-Salette, Quebec.

In August 1905, a large landslide involving glaciolacustrine silt took place on the right bank of the Thompson River at Spences Bridge (Fig. 2b), in southwestern British Columbia. The landslide rapidly entered the river and crossed it. Debris covered part of a First Nation settlement. The landslide generated a displacement wave in the river that swept over parts of the settlement, ran up the opposite valley wall to a height of 22.5 m above the river, and then flowed back into the channel. Calculations based on this runup and using the simple energy head formula, $v=(2gh)^{-\frac{1}{2}}$, indicate a wave velocity in the order of 21 m/s (75 km/h; Fig. 1). A wave 4.5 to 6 m high also rushed upstream, carrying debris over 1.5 km. Twenty buildings in the settlement, many of them log houses, were destroyed and fifteen people were killed including the elderly Chief Lillooet.

A similar event took place in western Quebec in 1908, when a landslide on the Lièvre River at Notre-Dame-de-la-Salette (Fig. 2b) resulted in the deaths of 33 people. At about 04:30 h on 26 April, a landslide occurred suddenly on the west bank of the Lièvre River, which at the time was frozen and had about 0.5 m of ice on its surface. Three houses on the west bank were immediately engulfed in the landslide, resulting in six deaths. On entering the river, the landslide material generated a highly destructive displacement wave that overwhelmed part of the village of Notre-Dame-de-la-Salette, located at that time on a low terrace opposite the landslide site. An additional 27 people were killed. Large blocks of river ice contained in the wave contributed to its remarkable destructive effect. Twelve wooden buildings were completely destroyed by the wave (Fig. 41), including a large, two-storey building housing the post office, general store, blacksmith's shop, and village inn. Debris was found up to 15 m above river level. Calculations based on this runup indicate a wave velocity in the order of 17 m/s (61 km/h; Fig. 1).

Underwater landslides may also cause landslide-generated waves, particularly in the confined fiords and bays of Canada's Pacific Coast. In 1975, for example, at Kitimat (Fig. 2b), in northwestern British Columbia, a large failure on the underwater slope of the Kitimat River delta, described above, generated an 8 m wave that swept around the shores of Kitimat Arm, damaging wharf facilities and boats at their moorings. In Eastern Canada, the 1929 earthquake-triggered submarine landslide off the Grand Banks, described above, generated a deadly tsunami that swept ashore 133 minutes later on the Burin Peninsula, Newfoundland (Fig. 2b), approximately 250 km to the north. The tsunami coincided with a storm high tide and was 4 to 12 m in height when it struck the coast. It took 27 lives and caused considerable property loss (*see* Clague, 2001).

Figure 41.

Contemporary photograph of destruction at Notre-Dame-de-la-Salette, western Quebec, showing the effects of a landslide-generated wave on wooden buildings on the east bank of the Lièvre River in April 1908. The view is upriver. The vertical arrow points to the landslide scarp on the west bank of the river. Photograph courtesy of the National Archives of Quebec

Landslide dams

Landslide dams cause damage upstream due to backflooding and downstream due to outburst floods that may develop as a result of a sudden break of the dam (*see* Brooks et al., 2001). Large landslides in glaciolacustrine sediments along the Thompson River, near Ashcroft, British Columbia (Fig. 2b), for example, have blocked the river on several occasions since the middle of the nineteenth century; in 1880, a large landslide blocked the river for 44 hours, forming a lake that extended 14 km upstream. Landslides in leda clay have also formed significant landslide dams that have caused both upstream and downstream flooding during the historical period.

LANDSLIDE DISASTERS IN CANADA

Recent work defined a landslide disaster in the Canadian context as a landslide event, or related geotechnical failure, that directly or indirectly results in at least three deaths. This criterion is largely arbitrary, but seems to approximate a Canadian disaster-reporting threshold. It is noted that for this paper, the definition does not include a consideration of the costs of material damage.

A total of 43 events that meet the nominal national landslide disaster criterion defined above occurred in Canada in the 159 years between 1840 and 1998. (For the purposes of a Canadian landslide disaster listing, the historical period is taken to begin in 1840. Newspapers were well established by that time and the year marks the occurrence of the first Canadian landslide that was subject to a scientific report.) This is equivalent to a landslide disaster frequency of one every 3.7 years, or an annual frequency of 0.27. These disasters resulted in at least 570 deaths (*see* Fig. 42, in pocket).

Landslide disasters are heavily concentrated in British Columbia (21 events; 49% of the total) and Quebec (17 events; 39.5% of the total), which together account for 88.5% of Canada's landslide disasters. The remainder occurred in Ontario (two events; 4.7%), Newfoundland (two events; 4.7%) and Alberta (one event; 2.3%). Of landslide-related deaths, 239 (42%) have occurred in British Columbia and 213 (37%) in Quebec. In the province of Quebec, 93 deaths (44%) were due to the four rockslides that occurred on Champlain Street, in Québec (Fig. 2b, 43). The 1903 Frank Slide (Fig. 2b, 44) is Alberta's only landslide disaster, but it is Canada's worst. Of the landslide-related

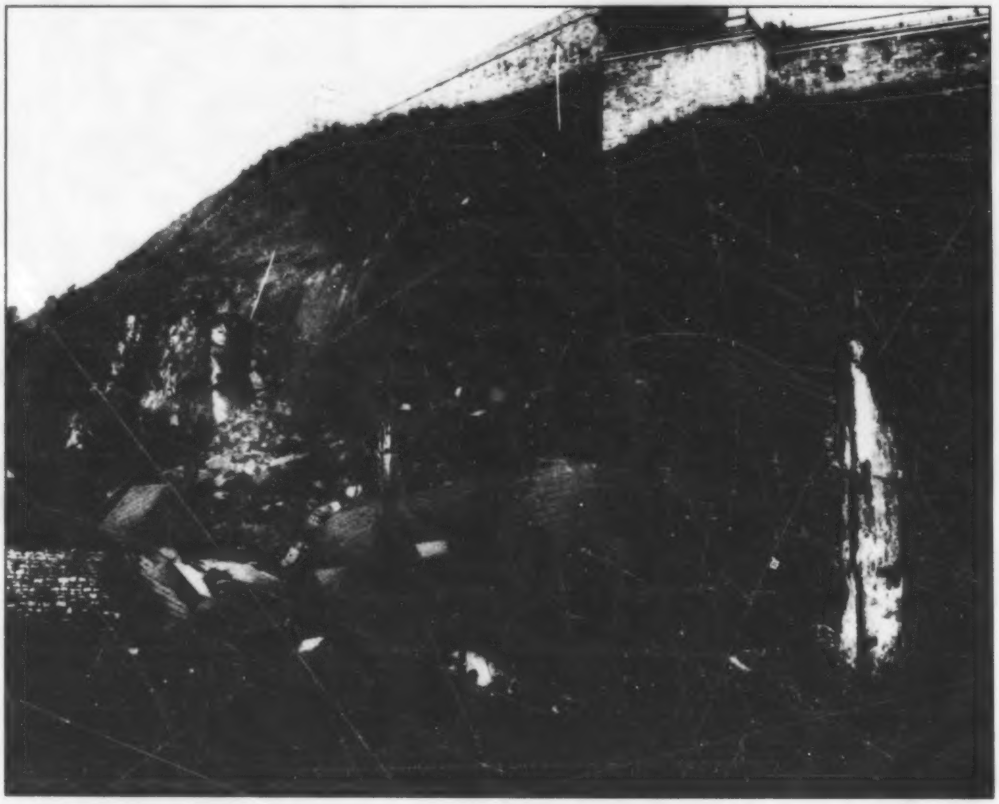
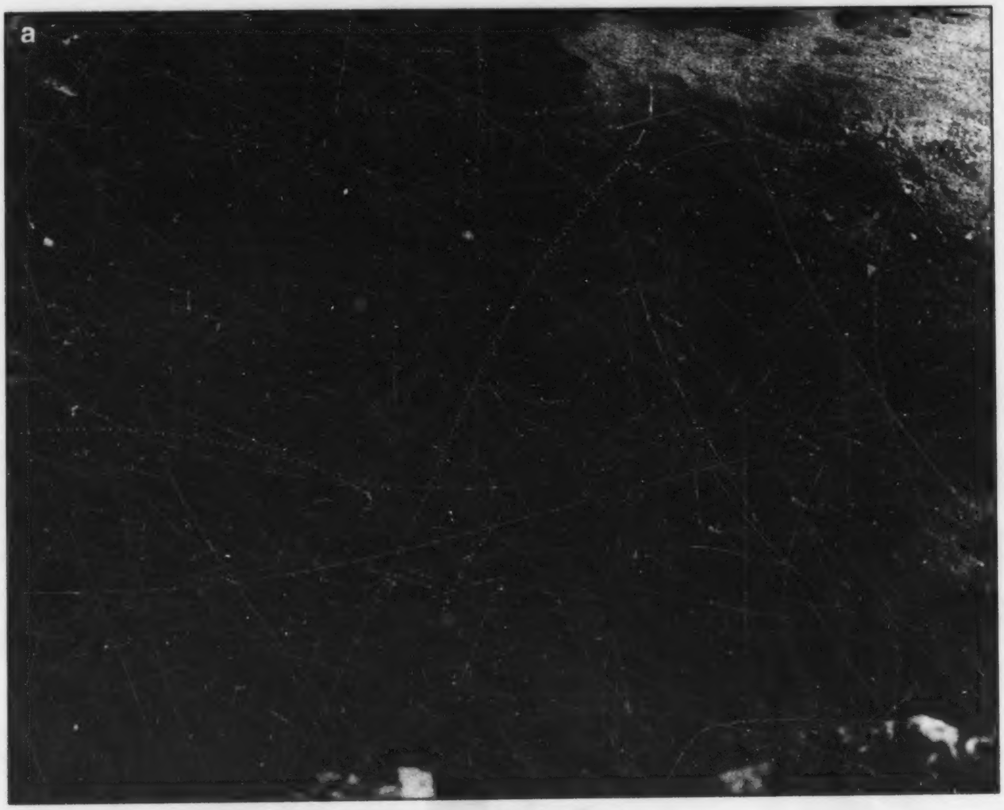


***Figure 43.** Contemporary photograph of the 1889 Champlain Street landslide disaster, Québec, Quebec. The view is to the southwest part of the landslide. Fifty people were crushed to death in the houses at the base of the failed slope. Note the walls of the Citadel at the top of the rock slope. Public Archives of Canada 131073*

deaths in Canada, 55% occurred in the Cordillera of Western Canada and 32.6%, in the St. Lawrence Lowlands of Eastern Canada.

Based on data in Figure 42, the most destructive landslide type in Canada was small-scale rockslides and rockfalls involving volumes of less than 100 000 m^3. These caused 27% (155) of the deaths in seven events across Canada. Second, in terms of destructiveness, were landslides in leda clay which caused 17.0% (98) of the deaths in nine events in the St. Lawrence Lowlands of Quebec. The third most destructive type was geotechnical failures involving the failure of man-made slopes, which accounted for 84 deaths, 15% of the total. The fourth most destructive landslide type consists of rock avalanches involving volumes in excess of 10 x 10^6 m^3. Rock avalanches caused 15% (83) of the total deaths in only three disaster events in the Cordillera. Fifth were debris flows and debris avalanches involving volumes of 60 000 m^3 or less. These landslides caused 11% (65) of the disaster deaths in 13 events. One hundred and twenty-five deaths (22% of the total) were caused by rapid flowslide-type movements in Quaternary sediments (glaciolacustrine and glaciomarine) in



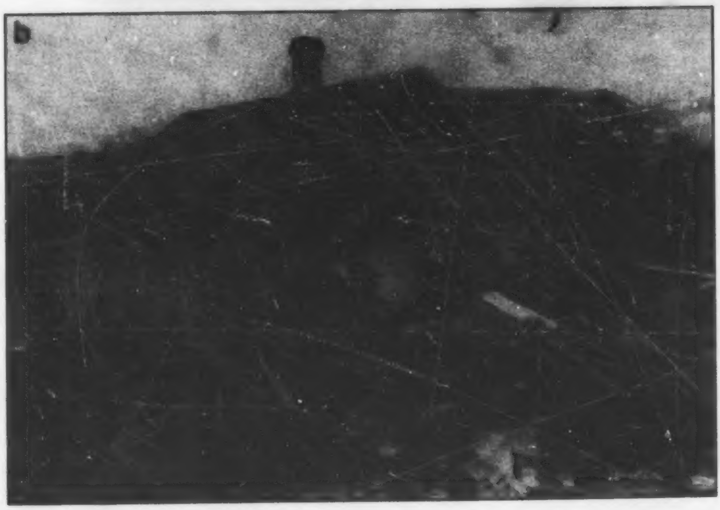


Figure 44.

a) View over the town of Frank, Alberta, and Frank Slide debris from the north shoulder of Turtle Mountain in 1911. Photograph by G.S. Rice. Public Archives of Canada PA 45409
b) One of the seven cottages destroyed by the Frank Slide on the outskirts of Frank, Alberta. The occupants of this house were asleep when the slide struck and were fortunate to survive. Glenbow Archives photograph NA-586-2

12 events in British Columbia and Quebec. No simple correlation exists between deaths per event and event magnitude, nor is the record dominated by large-magnitude events usually associated with landslide disasters.

Many of the landslide-related deaths were caused by the secondary effects of landslides and related geotechnical failure, i.e. landslide-generated displacement waves and outburst floods. Five of these events have caused 26% (147) of all landslide-related deaths.

Landslide disasters by their very nature occur in areas where human activity commonly frequently has altered the natural condition of the failed slope. Changes in slope condition due to deforestation, irrigation, and surface or subsurface excavation are inherent in development. Fills or embankments may load slopes and block surface or subsurface drainage. Blasting may result in the dynamic loading of slopes. Thus, a listing of historical landslide disasters includes events that are partly, or possibly largely, due to some element of human forcing. One hundred and thirty (23%) of landslide-related deaths that occurred in Canada from 1840 to 1998 were due to landslides caused directly or indirectly by human activity. These include deaths due to embankment failures, a landslide in the surface that penetrated underground mine workings, and two leda clay landslides caused by blasting. If we include the problematical 1903 Frank Slide in Alberta, which is thought to have been triggered by the effects of underground coal mining at the base of Turtle Mountain, this number rises to 205, or 36% of the total deaths, i.e. just over one in three.

Landslide disasters also result from human negligence. An incredible series of disasters on Champlain Street in Québec, Quebec, occurred between 1841 and 1889. In four landslide events, a total of 91 people died in rockslides that originated in the steep cliffs of Cape Diamant above Champlain Street. In returning its verdict after the 1889 disaster (Fig. 43), a coroner's jury found that the deaths were due "...to the gross and culpable negligence of the Federal authorities of the Dominion of Canada, in not taking the necessary and timely precautions...." (*Quebec Daily Mercury*, September 30, 1889, p. 3), which would have involved removing dwellings from the foot of the cliffs of Cape Diamant after the earlier disasters.

In the case of the 1921 Britannia Beach outburst flood (Fig. 2b) in southwestern British Columbia, which resulted in 56 deaths, a coroner's jury found that the disaster was caused by the collapse of culverted railway fill that had become blocked, thus damming the rain-swollen Britannia Creek. The disaster was thus caused by a geotechnical failure and the jury declared that "...it was criminal neglect on the part of the Britannia Mining and Smelting Co. Ltd...for deliberately allowing the blocking of...Britannia Creek causing a menace to persons living at Britannia Beach." (*Vancouver Daily Province*, November 29, 1921, p. 17).

Thus, human activity has interacted with natural systems in complex ways so that in many cases, it is probably impossible to distinguish between natural and human-caused disasters.

LANDSLIDES AND THE ENVIRONMENT

The balance of forces that control slope stability is very sensitive to changes in slope environment due to climatic, geophysical and anthropogenic factors. Landslides are a response to environmental change. They also have an impact on the environment by affecting the quality of primary resources such as forests, fish, and water supply. Landslides may also have an impact on other geomorphic systems. Landslide dams, for example, may affect river systems leading to aggradation downstream from the landslide debris.

Hydrometeorological and seismic triggers

Since water pressures are a key component in slope stability, hydrometeorological factors are important in controlling slope movements. Many types of landslides occur in response to heavy rainfall. The correlation is most marked in shallow landslides such as debris avalanches, debris flows, and slides in man-made embankments. In addition, large-scale, deep-seated slides may show a movement spike in response to a rainfall-induced water-pressure pulse. Rain-on-snow events (*see* Brooks et al., 2001) are particularly problematical with respect to slope stability because of the additional water generated in the slope water system by snowmelt. The threshold for rainfall-triggered landslides is highly variable because of such factors as antecedent slope water conditions and debris availability. Rockfalls can be caused by heavy rains, although they are more frequently due to freeze-thaw processes. Spring snowmelt is also a major landslide trigger; most landslides in leda clay, for example, occur in April.

Earthquakes are also important landslide triggers (Fig. 7, 45). Historical earthquakes have triggered a range of natural slope failures in both Eastern and Western Canada as well as numerous failures in road and railway embankments. The 1946 Vancouver Island earthquake (M7.3), for example, triggered many rockfalls and several rock avalanches as well as underwater landslides. The earthquake triggered a rock avalanche from Mount Colonel Foster (Fig. 2b) that entered Landslide Lake (Fig. 45). The rapidly moving rock-avalanche debris displaced lake water that swept downstream along the valley bottom, destroying mature forest up to 3 km from the lake (Fig. 45). More recently, several major rock avalanches were triggered in the St. Elias Mountains by the 1979 St. Elias earthquake (M7.5). In the 1988 Saguenay earthquake (M5.9), seismic shaking triggered rockfalls and rockslides in the Laurentian Mountains of Quebec and more than 20 failures occurred in natural slopes and embankments.

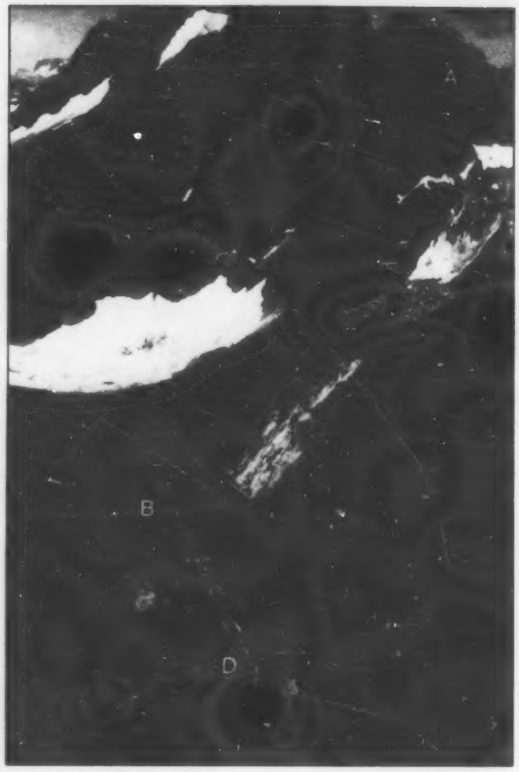
In 1663, Eastern Canada (then called 'Nouvelle-France') was struck by a major earthquake (M ~7.0). Although the region was sparsely populated (the European population of Nouvelle-France in 1665 was 3215), the direct effects of the earthquake are recorded in correspondence by members of religious orders active in the area at the time. Indirect evidence of landslides caused by the earthquake have been obtained by radiocarbon dating of landslide debris, dendrochronology, and the analysis of sediments in lakes and fiords. It is evident from these studies that widespread landsliding took place throughout Nouvelle-France.

including earthquake-triggered, large-scale, submarine failures of sediments in the Saguenay Fiord and massive landsliding along the Saint-Maurice River near Shawinigan (Fig. 46). A gigantic landslide (estimated volume 188 x 10^6 m^3) on the north side of the Saguenay River, within which the 1971 Saint-Jean-Vianney landslide took place, is thought to have been triggered by the 1663 earthquake. The debris from this landslide may have temporarily blocked the Saguenay River.

Logging, urbanization, mining, and irrigation

Human modification of the land surface (including its water balance) is a major factor in slope instability. In turn, landslides linked to human activity such as logging, urbanization, mining, and irrigation are major contributors to the impacts of these activities on the environment.

Figure 45. *Oblique aerial photograph taken in 1984 of Mount Colonel Foster, British Columbia, showing the source of the 1946 rock avalanche (A) and its path to Landslide Lake (B). The landslide was triggered by the 1946 Vancouver Island earthquake. Note the part of the debris that was deposited on a ledge above the lake (C) and the swath cut by a landslide-generated wave as it entered the upper reaches of the Elk River valley (D). Photograph by S.G. Evans. GSC 2000-099*



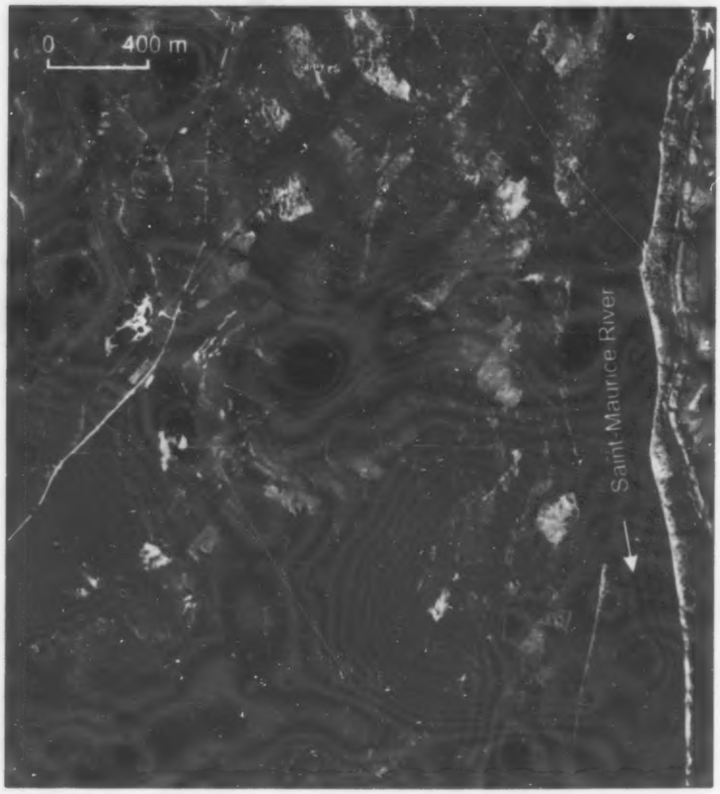


Figure 46.

Landslides in leda clay along the west side of the Saint-Maurice River, southwest of Shawinigan, Quebec, which radiocarbon dating suggests were triggered by the 1663 St. Lawrence earthquake. The black arrows define a line of landslide scarps. Province of Quebec Q64550-035

The effects of logging on slope stability are especially prevalent in the Cordillera (Fig. 47) where debris flows and debris avalanches have adversely affected Pacific salmon spawning grounds, removed productive land from forestry use, and contaminated water supplies. As newspaper headlines indicate (some examples are "Massive landslides in logging areas alarm habitat specialists (Vancouver Sun, 4 December 1995); "Clayoquot slide a record: clearcut logging cited in 'disaster'" (The Province, 23 January 1996)), the relationship between forestry practices and the occurrence of debris flows and debris avalanches became a major public issue in British Columbia and was instrumental in the development of British Columbia's Forest Practices Code, enacted in 1994. The Code is designed to mitigate the impact of logging practices on terrain stability, thus reducing the loss of productive forest lands, and to minimize damage to fish and wildlife habitat. In the Code, a careful evaluation of landslide hazards in any proposed forest development area is seen as critical to good forest resource management.

Urbanization modifies surface runoff, reduces water losses through evapotranspiration, and may add significant amounts of water to natural groundwater systems through lawn and garden irrigation. For example, the urbanization of the Calgary landscape has changed the slope microclimate from semi-arid to subhumid. This has led to higher groundwater levels and an increase in first-time slides in marginally stable slopes and to the reactivation of pre-existing slope failures.

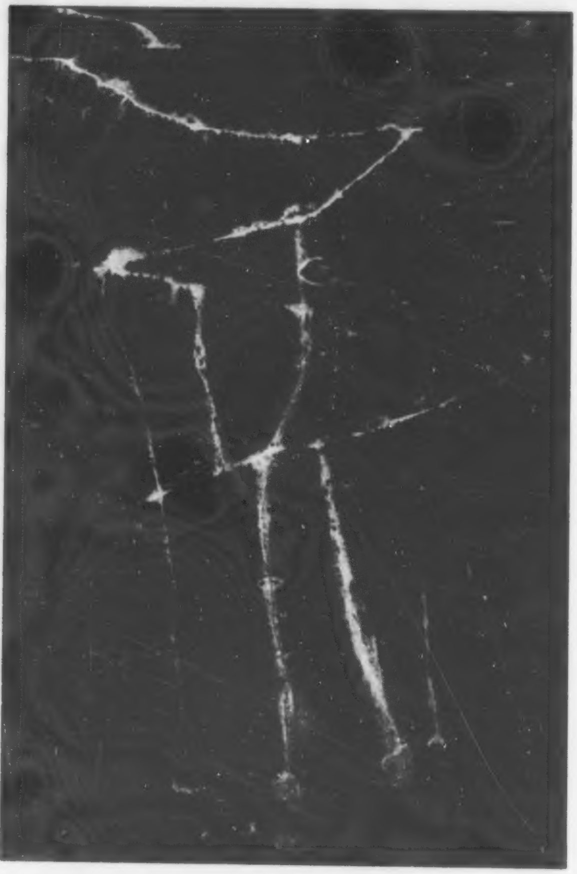


Figure 47. Debris avalanches and debris flows associated with logging, Cascade Mountains, near Chilliwack, southern British Columbia. Photograph by S.G. Evans. GSC 2000-100

Mining modifies stresses in a slope, which may lead to instability as was the case at Asbestos, Quebec, described above. With respect to underground mining, recent research on the 1903 Frank Slide suggests that Canada's largest landslide disaster resulted from displacements induced by coal mining at the foot of Turtle Mountain.

In drier parts of Canada, agriculture is only viable with irrigation. This significantly increases the infiltration of water into a slope system, leading to the destruction of soil suction and an increase in water pressures. Irrigation has been associated with some major destructive failures, particularly in the Pleistocene silt deposits of British Columbia's semi-arid, southern Interior Plateau (e.g. the 1905 Spences Bridge landslide referred to above).

Landslides and the Pacific salmon

In one of the wonders of Nature, Pacific salmon return to their original spawning grounds after spending three to five years in the open ocean. This upstream migration is a major part of the life cycle of the fish and its success is critical to the health of the Pacific salmon fishery in British Columbia and neighbouring parts of the United States. Major landslides have occurred in river systems draining into the Pacific in both prehistoric and historic times and have blocked or obstructed river flow, forming a barrier to upstream salmon migration. The barrier may involve a physical blockage of the river channel or the formation of flow conditions that make it impossible for salmon to pass. On the Fraser River, a rock avalanche downstream of Lillooet blocked the river, preventing the upstream migration of salmon. This event is thought to have caused a severe drop in salmon numbers, which resulted in food shortages for First Nation peoples in the Lillooet area who were dependent on the fish, leading to the abandonment of their villages at about AD 800.

In historic times, landslide dams have formed several times along the Thompson and Chilcotin rivers, tributaries of the Fraser River. Because the landslide debris consisted of easily erodible silt, these landslide dams were very short lived and did not present long-term problems for salmon migration. However, the infamous blockages and obstructions of the Fraser River at Hell's Gate and at a number of other locations in the Fraser Canyon by rockfall and construction debris during railway construction in 1913–1914 decimated salmon stocks, which have not yet fully recovered. Although these obstructions were only partly due to landslides, they illustrate how sensitive salmon stocks are to disturbances in flow patterns along migratory routes. Mitigation techniques have been adopted at substantial cost and include the direct removal of landslide debris from the river channel, river training works, and the construction of fishways to bypass the obstruction (Fig. 48). Landslides may also threaten the

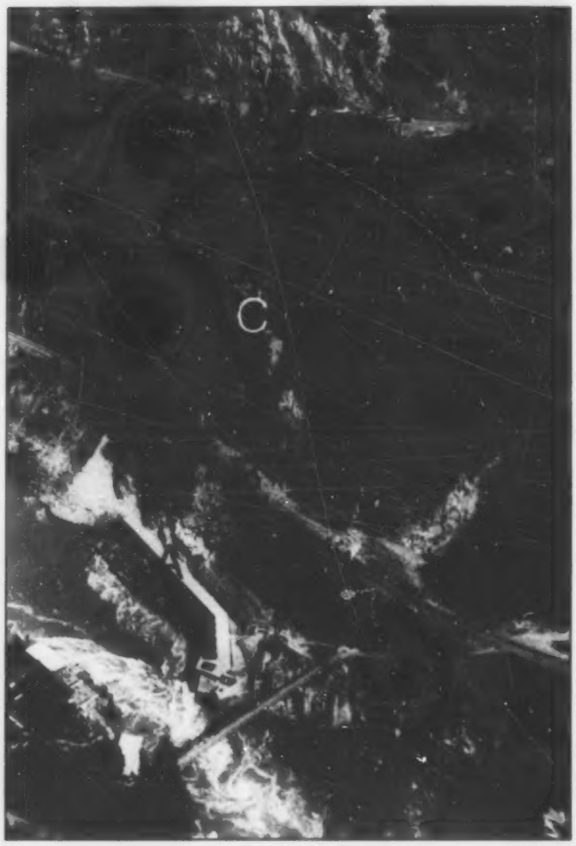


Figure 48. Aerial view of Hell's Gate, Fraser River, southwestern British Columbia. Note the fishways (A) built to assist in the migration of Pacific salmon upstream, the rockfall shed and gallery (B) constructed to protect the Canadian National Railway track, and the scar (C) of the 1914 rockfall. The Trans-Canada Highway is visible at the top of the photograph. Photograph by S.G. Evans. GSC 2000-101

spawning habitat by introducing sediment into critical river reaches, thus increasing the siltation of gravel spawning grounds.

Climate change

Elements of climate change currently being projected by global climatic models include an increase in precipitation, in the frequency of extreme climatic events, and in temperature. Landslides are heavily influenced by these climatic factors and their occurrence is therefore highly sensitive to climate change.

Under conditions of possible increased precipitation, future climatic change will result in an increase in the frequency of debris flows and other types of landslides. As in the past, these events will be expected to have an impact on settlements, infrastructural elements, resources, and the environment, resulting in human and financial losses.

Long-term temperature change affects the volume of glacier ice in mountain regions. Glacier-ice loss (includes glacier thinning, i.e. the decrease in elevation of the glacier surface, and glacier retreat, i.e. the recession of the snout of the glacier) due to global warming has been identified as an important factor in the occurrence of mountain-slope movements and rock avalanches. With respect to predicted temperature increases, further glacier-ice losses will result in continued debutressing (debutressing of slopes due to glacier thinning removes supportive weight from the slope and may lead to instability) of mountain slopes leading to slope deformation (Fig. 49) and, in some cases, catastrophic failure. The potential impact of rock avalanches should therefore be considered in the development of areas adjacent to and downstream from present-day glaciers. Increased temperature will also affect permafrost slopes. Debris flows have been associated with the decay of mountain permafrost during recent

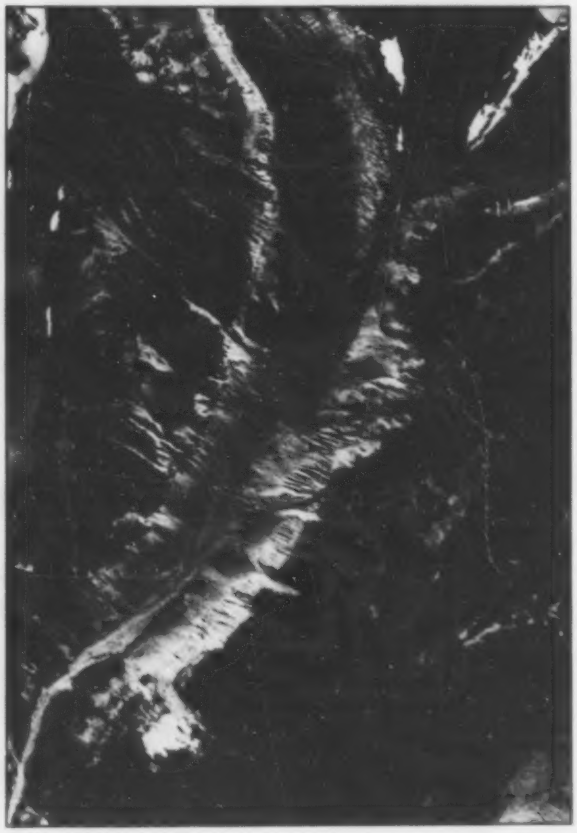


Figure 49. Vertical aerial photograph of extensive slope deformation at Affliction Creek, southern Coast Mountains, British Columbia. The arrows point to tension cracks and disturbed ground. Movement results from debutressing of the slope due to the retreat and downwasting of Affliction Glacier (A) since it reached its Neoglacial maximum extent at about AD 1840. Note the marked Neoglacial trimline in Affliction Creek. Photograph taken September 1973. Province of British Columbia BC7551; 218

warming. In northern permafrost areas, the further decay of permafrost as a result of continued warming is likely to increase the occurrence of thaw flowslides and other types of landslides. Locally, forest fires will amplify this effect. The mitigation of landslide hazard is a vital component of adaptation strategies to minimize the impact of future climate change.

REFERENCES

Anonymous
1894: The Saint-Alban landslide near Québec; The Railroad Gazette, v. 26, p. 458–460
Brooks, G.R., Evans, S.G., and Clague, J.J.
2001: Floods; *in* A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548.
Clague, J.J.
2001: Tsunamis; *in* A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548.
CN North America
1993: Eye on Safety, v. 3, p. 9.
Enegren, E.G. and Moore, D.P.
1992: Branham Ridge Slide — twenty-two years of slope monitoring; *in* Landslides, (ed.) D.H. Bell; Proceedings, Sixth International Symposium on Landslides, v. 2, p. 1129–1134.
Kelly, A.J., Sauer, E.K., Christiansen, E.A., Barbour, S.L., and Widger, R.A.
1995: Deformation of the Deer Creek Bridge by an active landslide in clay shale; Canadian Geotechnical Journal, v. 32, p. 701–724.
Morgenstern, N.R., Fair, A.E., and McRoberts, E.C.
1988: Geotechnical engineering beyond soil mechanics — a case study; Canadian Geotechnical Journal, v. 25, p. 637–661.
Plafker, G. And Ericksen, G.E.
1978: Nevados Huascarán avalanches, Peru; *in* Rockslides and Avalanches, 1, Natural Phenomena, (ed.) B. Voight; Elsevier Scientific Publishing Company, New York, p. 277–314.
Prior, D.B., Bornhold, B.D., and Johns, M.W.
1984: Depositional characteristics of a submarine debris flow; Journal of Geology, v. 92, p. 707–727.
Scott, J.S. and Brooker, E.E.
1968: Geological and engineering aspects of Upper Cretaceous shales in western Canada; Geological Survey of Canada, Paper 66-37, 75 p.
Smith, S.L., Burgess, M.M., and Heginbottom, J.A.
2001: Permafrost in Canada, a challenge to northern development; *in* A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548.
Thomson, S. and Hayley, D.
1975: The Little Smoky Landslide; Canadian Geotechnical Journal, v. 12, p. 379–392.

SUGGESTED READING

Landslide overviews, landslide classification, landslide mechanics and stabilization techniques

Dikau, R., Brunsden, D., Schrott, L., and Ibsen, M-L. (ed.)
1996: Landslide Recognition; John Wiley and Sons, New York, 251 p.
Hutchinson, J.N.
1988: Morphological and technical parameters of landslides in relation to geology and hydrogeology, state-of-the-art report; *in* Landslides, (ed.) C. Bonnard; Proceedings, Fifth International Symposium on Landslides, v.1, p. 3–35.
Turner, A.K. and Schuster, R.L. (ed.)
1996: Landslides, investigation and mitigation, (ed.) A.K. Turner and R.L. Schuster; Transportation Research Board Special Report 247, National Academy Press, Washington, D.C., 673 p.

Overviews of landslides in Canada

Cruden, D.M., Bornhold, B.D., Chagnon, J-Y., Evans, S.G., Heginbottom, J.A., Locat, J., Moran, K., Piper, D.J.W., Powell, R., Prior, D., Quigley, R.M., and Thomson, S.
1989: Landslides: extent and economic significance in Canada; *in* Landslides: Extent and Economic Significance, (ed.) E.E. Brabb and B.L. Harrod; Balkema, Rotterdam, p.1–24.
Mollard, J.D.
1977: Regional landslide types in Canada; *in* Geological Society of America Reviews in Engineering Geology, v. 3, p. 29–56.

Rapid landslides in rock slopes

Ballivy, G., Hardy, J., LeBuis, J., and Bergeron, R.
1984: Techniques d'abattage d'une paroi rocheuse instable; *in* Proceedings, Fourth International Symposium on Landslides, Toronto, Ontario, v. 1, p. 455–459.
Benko, B. and Stead, D.
1998: The Frank slide: a reexamination of the failure mechanism; Canadian Geotechnical Journal, v. 35, p. 399–311.
Boyd, R.D.
1991: Rockfall containment measures at Springdale, Newfoundland; *in* Slope Stability Engineering; Developments and Applications, (ed.) R.J. Chandler; Thomas Telford Ltd., London, England, p. 409–413.
Eisbacher, G.H.
1979: Cliff collapse and rock avalanches (sturzstroms) in the Mackenzie Mountains, northwestern Canada; Canadian Geotechnical Journal, v. 16, p. 309–334.
Evans, S.G. and Brooks, G.R.
1991: Prehistoric debris avalanches from Mount Cayley volcano, British Columbia; Canadian Journal of Earth Sciences, v. 28, p. 1365–1374.
Evans, S.G. and Hungr, O.
1993: The analysis of rockfall hazard at the base of talus slopes; Canadian Geotechnial Journal, v. 30, p. 620–636.
Evans, S.G., Aitken, J.D., Wetmiller, R.J., and Horner, R.B.
1987: A rock avalanche triggered by the October 1985 North Nahanni earthquake, District of Mackenzie, N.W.T.; Canadian Journal of Earth Sciences, v. 24, p. 176–184.
Evans, S.G., Clague, J.J., Woodsworth, G.J., and Hungr, O.
1989: The Pandemonium Creek rock avalanche, British Columbia; Canadian Geotechnical Journal, v. 26, p. 427–446.
Evans, S.G., Hungr, O., and Enegren, E.G.
1994: The Avalanche Lake rock avalanche, Mackenzie Mountains, Northwest Territories, Canada; description, dating, and dynamics; Canadian Geotechnical Journal, v. 31, p. 749–768.
Kerr, J.W.
1990: Frank Slide; Barker Publishing Ltd., Priddis, Alberta, 49 p.
Mathews, W.H. and McTaggart, K.C.
1978: Hope rockslides, British Columbia, Canada; *in* Rockslides and Avalanches, 1, Natural Phenomena, (ed.) B. Voight; Elsevier Scientific Publishing Company, New York, p. 259–275.
McConnell, R.G. and Brock, R.W.
1904: Report on the great landslide at Frank, Alberta; Annual Report of the Department of Interior for the year 1902–1903, Part VIII, 17 p.
Moore, D.P. and Mathews, W.H.
1978: The Rubble Creek landslide, southwestern British Columbia; Canadian Journal of Earth Sciences, v. 15, p. 1039–1052.

Debris avalanches and channelized debris flows

Batterson, M., Liverman, D.G.E., Ryan, J., and Taylor, D.
1999: The assessment of geological hazards and disasters in Newfoundland: an update; *in* Current Research (1999); Newfoundland Department of Mines and Energy Geological Survey, Report 99-1, p. 95–123.
Clague, J.J., Evans, S.G., and Blown, I.G.
1985: A debris flow triggered by the breaching of a moraine-dammed lake, Klattasine Creek, British Columbia; Canadian Journal of Earth Sciences, v. 22, p. 1492–1502.

Hungr, O., Morgan, G.C., VanDine, D.F., and Lister, D.R.
1987: Debris flow defences in British Columbia; *in* Debris Flows/Avalanches: Process, Recognition, and Mitigation, (ed.) J.E. Costa and G.F. Wieczorek; Geological Society of America, Reviews in Engineering Geology, v. VII, p. 201–222.

Jackson, L.E., Jr.
1987: Debris flow hazard in the Canadian Rocky Mountains; Geological Survey of Canada, Paper 86-11, 20 p.

Lister, D.R., Kerr, J.W.G., Morgan, G.C., and VanDine, D.F.
1984: Debris torrents along Howe Sound, British Columbia; *in* Proceedings, Fourth International Symposium on Landslides, v. 1, p. 649–654.

VanDine, D.F.
1985: Debris flows and debris torrents in the southern Canadian Cordillera; Canadian Geotechnical Journal, v. 22, p. 44–68.
1996: Debris flow control structures for forest engineering. Research Branch, British Columbia Ministry of Forests, Victoria, British Columbia, Working Paper 22/1996, 68 p.

Slow slides

Baracos, A. and Graham, J.
1981: Landslide problems in Winnipeg; Canadian Geotechnical Journal, v. 18, p. 390–401.

Bovis, M.J. and Evans, S.G.
1996: Extensive deformations of rock slopes in southern Coast Mountains, southwest British Columbia, Canada; Engineering Geology, v. 44, p. 163–182.

Cruden, D.M., Keegan, T.R., and Thomson, S.
1993: The landslide dam on the Saddle River near Rycroft, Alberta; Canadian Geotechnical Journal, v. 30, p. 1003–1015.

Evans, S.G.
1982: Landslides and surficial deposits in urban areas of British Columbia: a review; Canadian Geotechnical Journal, v. 19, p. 269–288.

Grant, D.G.
1974: Terrain studies of Cape Breton Island, Nova Scotia and of the northern Peninsula, Newfoundland; Geological Survey of Canada, Paper 74-1, p. 241–246.

Hungr, O., Gerath, R.F., and Morgan, G.C.
1984: Landslides of the Scatter River area, northeastern British Columbia; Preprint volume, 37th Canadian Geotechnical Conference, Toronto, p. 113–119.

Imrie, A.S., Moore, D.P., and Enegren, E.G.
1992: Performance and maintenance of the drainage system at Downie Slide; *in* Landslides, (ed.) D.H. Bell; Proceedings, Sixth International Symposium on Landslides, v. 1, p.751–757.

Thomson, S. and Mekechuk, J.
1982: A landslide in glacial lake clays in central British Columbia; Canadian Geotechnical Journal, v. 19, p. 296–306.

Thomson, S. and Morgenstern, N.R.
1979: Landslides in argillaceous rocks, Prairie Provinces, Canada; *in* Rockslides and Avalanches, v. 2, Engineering Sites, (ed.) B. Voight; Developments in Geotechnical Engineering 14B, Elsevier, Amsterdam, p. 515–540.

Slow earthflows in rocks and Quaternary sediments

Bovis, M.J.
1985: Earthflows in the Interior Plateau, southwest British Columbia; Canadian Journal of Earth Sciences, v. 22, p. 313–334.

VanDine, D.F
1980: Engineering geology and geotechnical study of Drynoch landslide, British Columbia; Geological Survey of Canada, Paper 79-31, 34 p.

Rapid slides and earthflows in Quaternary sediments

Dawson, G.M.
1899: Remarkable landslip in Portneuf County, Quebec; Bulletin of the Geological Society of America, v. 10, p. 484–490.

Eden, W.J. and Mitchell, R.J.
1971: South Nation River landslide, 16 May 1971; Canadian Geotechnical Journal, v. 8, p. 446–451.

Evans, S.G. and Brooks, G.R.
1994: An earthflow in sensitive Champlain Sea sediments at Lemieux, Ontario, June 20, 1993, and its impact on the South Nation River; Canadian Geotechnical Journal, v. 31, p. 384–394.

Evans, S.G., Hu, X-Q., and Enegren, E.G.
1996: The 1973 Attachie Slide, Peace River Valley, near Fort St. John. British Columbia, Canada: a landslide with a high velocity flowslide component, in Pleistocene sediments; *in* Proceedings, Seventh International Symposium on Landslides, Trondheim, Norway, v. 2, p. 715–720.

Evans, S.G., Begin, C., Lawrence, D.E., Demers, D., Grondin, G., Aylsworth, J.M., Parent, M., Michaud, Y., and Brooks, G.R.
1997: A massive retrogressive landslide in sensitive Champlain Sea sediments near St-Boniface-de-Shawinigan, Quebec, Canada, April 1996; Landslide News (Japan Landslide Society), v. 10, p. 17–19.

Geertsema, M. and Schwab, J.W.
1997: Retrogressive flowslides in the Terrace–Kitimat area, British Columbia: from early post-deglaciation to present and implications for future slides; *in* Forestry Geotechnique and Resource Engineering; Proceedings, 11th Annual Vancouver Geotechnical Society Symposium, p. 115–133.

Mitchell, R.J. and Markell, A.R.
1974: Flowsliding in sensitive soils; Canadian Geotechnical Journal, v. 11, p. 11–31.

Tavenas, F., Chagnon, J-Y., and LaRochelle, P.
1971: The Saint-Jean-Vianney landslide: observations and eyewitness accounts; Canadian Geotechnical Journal, v. 8, p. 463–478

Landslides in permafrost terrain

Aylsworth, J.M., Duk-Rodkin, A., Robertson, T., and Traynor, J.A.
in press: Landslides of the Mackenzie valley and adjacent mountainous and coastal regions; *in* The Physical Environment of the Mackenzie River Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547.

Broscoe, A.J. and Thomson, S.
1969: Observations on an Alpine mudflow, Steele Creek, Yukon; Canadian Journal of Earth Sciences, v. 6, p. 219–229.

Dyke, L.D.
in press: Stability of permafrost slopes in the Mackenzie valley; *in* The Physical Environment of the Mackenzie River Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547.

Harry, D.G. and MacInnes, K.L.
1988: The effect of forest fires on permafrost terrain stability, Little Chicago–Travaillant Lake area, Mackenzie Valley, N.W.T.; *in* Current Research, Part D; Geological Survey of Canada, Paper 88-1D, p. 91–94.

Lewkowicz, A.G.
1990: Morphology, frequency and magnitude of active-layer detachment slides, Fosheim Peninsula, Ellesmere Island, N.W.T.; *in* Permafrost – Canada, Proceedings, Fifth Canadian Permafrost Conference, (ed.) M.M. Burgess, D.G. Harry, and D.C. Sego; Université Laval, Québec, Quebec, p. 111–118.

Lewkowicz, A.G. and Harshorn, J.
1998: Terrestrial record of rapid mass movements in the Sawtooth Range, Ellesmere Island, Northwest Territories, Canada; Canadian Journal of Earth Sciences, v. 35, p. 55–64.

McRoberts, E.C.
1978: Slope stability in cold regions; *in* Geotechnical Engineering for Cold Regions, (ed.) O.B. Andersland and D.M. Anderson; McGraw-Hill, New York, p. 363–404.

Landslides in man-made dumps, embankments, and excavated slopes

Eden, W.J.
1964: Earthflows at the Beattie Mine, Quebec, Canada; Canadian Geotechnical Journal, v. 1, p. 104–114.

Gartshore, R.W.
1975: The history of overburden instability at Jeffrey Mine; Proceedings, Tenth Canadian Rock Mechanics Symposium, p. 1–26.

Hungr, O. and Kent, A.

1995: Coal mine waste dump failures in British Columbia, Canada; Landslide News (Japan Landslide Society), v. 9, p. 26–28.

Weichert, D.H., Horner, R.B., and Evans, S.G.

1994: Seismic signatures of landslides: the 1990 Brenda Mine collapse and the 1965 Hope rockslides; Bulletin of the Seismological Society of America, v. 84, p. 1523–1532.

Underwater landslides

Christian, H.A., Mosher, D.C., Mulder, T., Barrie, J.V., and Courtney, R.C.

1997: Geomorphology and potential instability on the Fraser River delta foreslope, Vancouver, British Columbia; Canadian Geotechnical Journal, v. 34, p. 432–446.

Piper, D.J.W., Shor, A.N., and Clarke, J.E.H.

1988: The 1929 "Grand Banks" earthquake, slump, and turbidity current; *in* Sedimentologic Consequences of Convulsive Geologic Events, (ed.) H.E. Clifton; Geological Society of America, Special Paper 229, p. 77–92.

Terzaghi, K.

1956: Varieties of submarine slope failures; Proceedings, Eighth Texas Conference on Soil Mechanics and Foundation Engineering, p. 1–41.

The secondary effects of landslides

Doxsee, W.W.

1948: The Grand Banks Earthquake of November 18, 1929; Publications of the Dominion Observatory, Ottawa, v. 7, no. 7, p. 323–335.

Evans, S.G.

1984: The 1880 landslide dam on Thompson River, near Ashcroft, British Columbia; *in* Current Research, Part A; Geological Survey of Canada, Paper 84-1A, p. 655–658.

1989: The 1946 Mount Colonel Foster rock avalanche and associated displacement wave, Vancouver Island, British Columbia; Canadian Geotechnical Journal, v. 26, p. 447–452.

Lapointe, P-L.

1974: Les éboulements dans la vallée de la Lièvre; Asticou, vol. 13, p. 23–33.

Landslide disasters in Canada (for Frank Slide see above)

Baillairgé, C.

1893: The Quebec land slide of September 19, 1889; Canadian Society of Civil Engineers, Transactions, v. 7, p. 140–152.

Evans, S.G.

1997: Fatal landslides and landslide risk in Canada; *in* Landslide Risk Assessment, (ed.) D.M. Cruden and R. Fell; A.A. Balkema, Rotterdam, p. 185–196.

1999: Landslide disasters in Canada 1840–1998; Geological Survey of Canada, Open File 3712.

Landslides and the environment

Church, M. and Miles, M.J.

1987: Meteorological antecedents to debris flow in southwestern British Columbia; some case histories; *in* Debris Flows/Avalanches: Process, Recognition, and Mitigation, (ed.) J.E. Costa and G.F. Wieczoreck; Geological Society of America, Reviews in Engineering Geology, v. VII, p. 63–79.

Desjardins, R.

1980: Tremblements de terre et glissements de terrain : corrélation entre des datations au ^{14}C et des données historiques à Shawinigan, Québec; Géographie physique et Quaternaire, vol. 34, p. 359–362.

Evans, S.G. and Clague, J.J.

1989: Rain-induced landslides in the Canadian Cordillera, July 1988; Geoscience Canada, v. 16, p. 193–200.

1994: Recent climatic change and catastrophic geomorphic processes in mountain environments; Geomorphology, v. 10, p. 107–128.

1997: The impact of climate change on catastrophic geomorphic processes in the mountains of Alberta, British Columbia, and the Yukon; *in* Responding to Global Climate Change in B.C. and the Yukon, (ed.) E. Taylor and W. Taylor; Environment Canada-British Columbia Ministry of Environment Lands and Parks, Victoria, British Columbia, p. 7-1 to 7-16.

Hardy, R.M., Clark, J.I., and Stepanek, M.

1980: A summary of case histories spanning 30 years of slope stabilisation in Calgary, Alberta; Proceedings, Specialty Conference on Slope Stability Problems in Urban Areas, Toronto; Canadian Geotechnical Society, 24 p.

Hayden, B. and Ryder, J.M.

1991: Prehistoric cultural collapse in the Lillooet area; American Antiquity, v. 56, p. 50–65.

Lefebvre, G., Lebouef, D., Hornych, P., and Tanguay, L.

1992: Slope failures associated with the 1988 Saguenay earthquake, Quebec, Canada; Canadian Geotechnical Journal, v. 29, p. 117–130.

Rogers, G.C.

1980: A documentation of soil failure during the British Columbia earthquake of 23 June, 1946; Canadian Geotechnical Journal, v. 17, p. 122–127.

Rood, K.M.

1984: An aerial photograph inventory of the frequency and yield of mass wasting on the Queen Charlotte Islands, British Columbia; British Columbia Ministry of Forests, Land Management Report no. 34, 55 p.

Schwab, J.W.

1983: Mass wasting: October–November 1978 storm, Rennell Sound, Queen Charlotte Islands, British Columbia; British Columbia Ministry of Forests, Research Note 91, 23 p.

Stanton, R.B.

1898: The great land-slides on the Canadian Pacific Railway in British Columbia (includes Discussion); Institution of Civil Engineers, Proceedings, v. 132, pt. 2, p. 1–46.

Syvitski, J.P.M. and Schafer, C.T.

1996: Evidence for an earthquake-triggered basin collapse in Saguenay Fjord, Canada; Sedimentary Geology, v. 104, p. 127–153.

Snow avalanches

Bruce Jamieson[1]

Jamieson, B. 2001: Snow avalanches; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 81–100.*

Abstract: Snow avalanches occur in many areas of Canada, mostly in remote or uninhabited locations. The few that impact society cause significant damage. Snow avalanches have killed over 570 people in Canada since the mid-1800s. Traffic delays and damage to forests cost millions of dollars annually.

Slab avalanches cause most of the hazard to people, property, and infrastructure. They are released by a failure within a weak layer or at a weak interface underlying one or more layers of cohesive snow. Point-release avalanches start in relatively cohesionless snow and are usually smaller.

Slope angle and orientation to wind strongly influence the formation of avalanches.

Avalanche hazards to buildings and transportation corridors can be mitigated using structural methods. Ski areas and transportation corridors often require forecasting and control programs to reduce the avalanche hazard.

About half the people buried in avalanches die within 30 minutes. Portable transceivers reduce search times and increase the odds of finding victims alive.

Résumé : Les avalanches de neige se produisent dans de nombreuses régions du Canada, principalement dans les zones éloignées ou inhabitées. Les quelques avalanches qui affectent notre société causent d'énormes dégâts. Elles ont causé la perte de 570 vies humaines au Canada depuis le milieu du XIX^{e} siècle. Les entraves à la circulation et les dommages causés aux forêts entraînent des coûts de l'ordre de millions de dollars chaque année.

Les avalanches en plaques constitue le plus grand danger pour les personnes, les biens et les infrastructures. Elles se déclenchent à la faveur d'une rupture dans une couche ou une interface mal soudée reposant sous une ou plusieurs couches de neige cohésive. Les avalanches à déclenchement ponctuel se produisent dans de la neige pulvérulente et sont généralement plus petites.

L'angle de la pente et l'orientation de celle-ci par rapport au vent influent fortement sur le déclenchement des avalanches.

Les dégâts que font subir les avalanches aux édifices et aux voies de communication peuvent être atténués en appliquant des méthodes structurelles. Les stations de ski et les voies de communication ont souvent besoin de programmes de prévision et de contrôle pour réduire les dégâts dus aux avalanches.

Presque la moitié des personnes ensevelies sous une avalanche meurent en 30 minutes. Les émetteurs-récepteurs portables réduisent la durée des recherches et augmentent les chances de trouver les victimes vivantes.

---

[1] Department of Geology and Geophysics, University of Calgary, Calgary, Alberta

INTRODUCTION

A snow avalanche is a mass of snow, usually more than several cubic metres in volume, that slides down a slope. Avalanches range from those too small to bury a person to large slides (up to 10^8 m^3) capable of destroying a village or many hectares of forest (Fig. 1).

Most avalanches occur in remote or uninhabited areas and do not affect people, property, or infrastructure. Approximately 1.5 million avalanches large enough to injure, bury, or kill a person occur annually in Western Canada. Yet only about 100 avalanche accidents (fatal, nonfatal, and property damage) are reported each year. Assuming that approximately one fifth of all avalanche accidents are reported, this implies that only about 1 in 3000 potentially destructive avalanches actually injures people or damages property and infrastructure.

SNOW AVALANCHES AND THEIR EFFECTS

Fatalities

Over 570 people have died in avalanches in Canada since the earliest reported accidents in the mid-1800s. Initially, many people were killed while building railways or roads or working at mine sites. However, the number of industrial and transportation fatalities has declined substantially in the twentieth century whereas the average number of recreational accidents each year has increased dramatically since the 1940s (Fig. 2). In Canada between 1993 and 1997, an average of 12 people died each year in snow avalanches; most were recreationists who were caught by an avalanche that they or a member of their group triggered.

Traffic delays

Snow avalanches that affect transportation corridors can cause significant traffic delays and economic losses. For example, based on a traffic volume of 350 vehicles an hour in

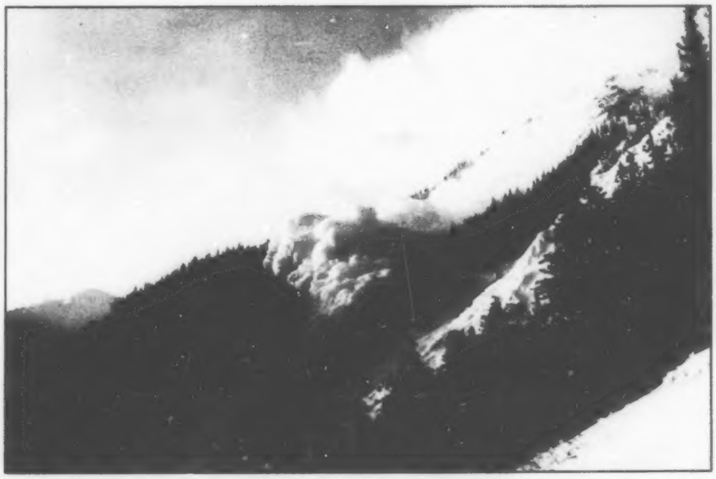
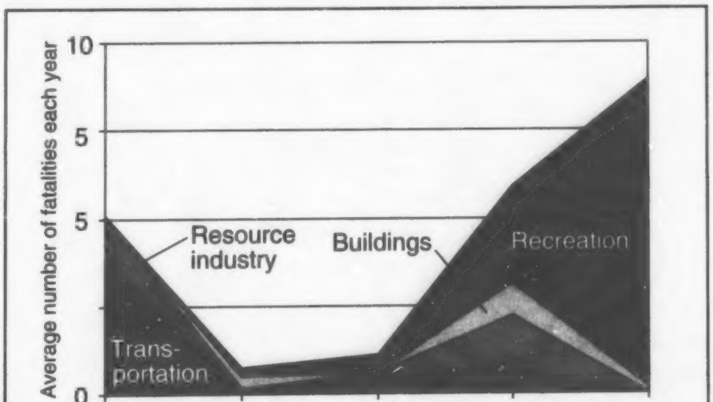


Figure 1.

A large snow avalanche with a powder cloud. Photograph courtesy of Parks Canada, ©Parks Canada #615-288-O-094

Figure 2.

Human activity during snow avalanche accidents categorized as recreational, transportation (including construction of transportation routes), resource industry (including accidents in work camps), and in or near commercial or residential buildings. A few accidents are excluded that do not fit any of these categories. Over time, the number of recreational avalanche accidents has increased dramatically whereas accidents related to transportation have decreased.



each direction, a two-hour closure at Rogers Pass, British Columbia, will cause monetary losses of $50 000 to $90 000, depending on the proportion of trucks in the delayed traffic. Since avalanche-related closures in Rogers Pass total about 100 hours in an average winter, the annual cost reaches a few million dollars. Cost estimates for closures in Three Valley Gap west of Revelstoke, British Columbia, and at Kootenay Pass in southern British Columbia are $700 000 and $400 000 annually. Considering the many other avalanche areas in the Cordillera managed by Parks Canada and the British Columbia Ministry of Transportation and Highways, the total annual cost of traffic delays exceeds $5 million. This figure is conservative as it does not include railway delays or the effect of closures on just-in-time delivery and the consequences of lost business.

Damage to property and forests

Reports of property damage caused by snow avalanches usually total less than $500 000 annually, although much damage probably goes unreported. This total includes, for example, damage to residential and commercial buildings, power lines, ski lifts, and vehicles. Damaged snowmobiles and lost skis are rarely reported, but also represent property damage.

Snow avalanches can damage forests by uprooting, breaking, or injuring trees (Fig. 3), including young trees planted in cutblocks. Avalanches can also remove organic soil, which is important for forest regrowth. Individual avalanches can damage up to $500 000 worth of timber. Forest harvesting may increase snow avalanche damage to timber because avalanches may start in logged areas or avalanches that start above the treeline may run through logged areas. Either type of avalanche may damage young trees and in places mature timber located farther downslope. In some cases, structures that support the snowpack can be installed to inhibit avalanches and thereby aid reforestation. However, this approach is expensive and not always successful.

AVALANCHE INITIATION

Loose snow or point-release avalanches start in relatively cohesionless snow. As they move downslope, adjacent snow is set in motion leaving a trough shaped like an inverted 'V' (Fig. 4). Although most point-release avalanches of dry snow are relatively small, point releases of wet snow are more likely to form larger, more destructive avalanches.

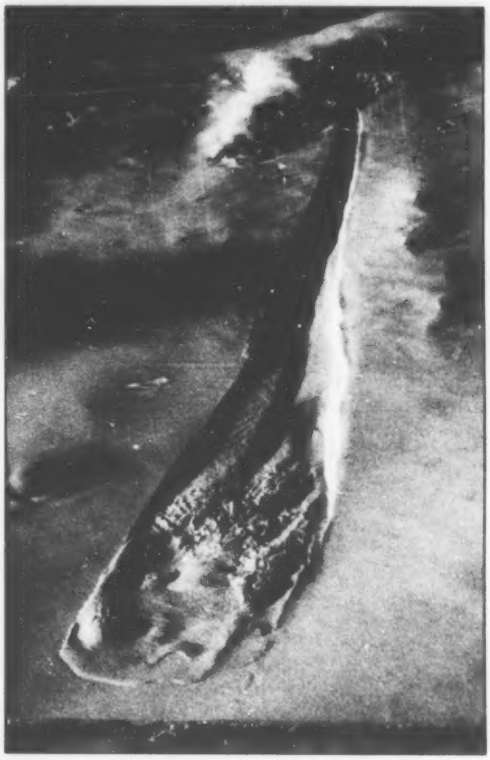


Figure 4. A point-release snow avalanche. Photograph by B. Jamieson. GSC 2000-102

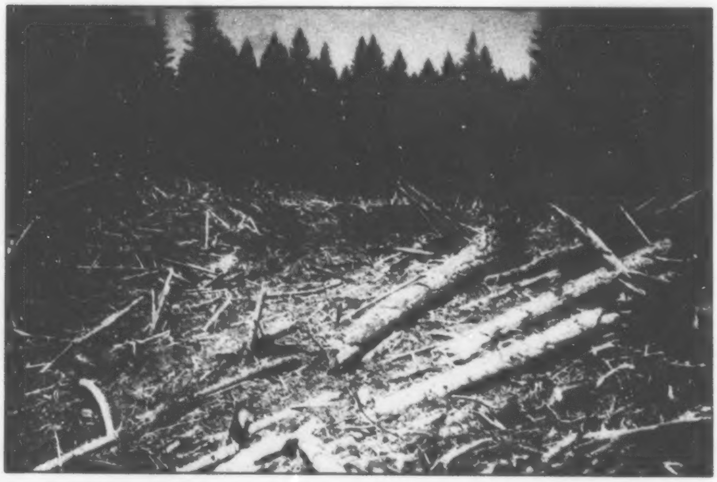


Figure 3.

Timber damaged by a snow avalanche. Photograph courtesy of Chris Stethem.

Alternatively, layers of cohesive snow can fail as a slab, leaving a distinct fracture line or crown fracture across the slope (Fig. 5). Slab avalanches are usually larger than point-release avalanches. Also, they are more hazardous to recreationists because they can be triggered from below the crown fracture or well below the start zone. For a slab to release, fracture must occur around the perimeter of the slab (crown, flanks, and toe or stauchwall) and at the base. Avalanche release begins when the failure spreads along a plane in a weak layer or weak interface (the failure plane) at the base of the slab (Fig. 6). As most dangerous avalanches are slab avalanches, backcountry recreationists and avalanche forecasters look for weak layers or weak interfaces in the snowpack when assessing the local avalanche hazard.

Weak layers

Weak snow layers that are sandwiched between stronger layers (Fig. 7) can form in different ways.

- If wind increases during a snowstorm, a dense layer of broken and packed snow crystals (the slab) may be deposited on top of a delicate layer of unbroken crystals (the weak layer) that fell before the wind increased.
- Weak layers of depth hoar can develop within the snowpack if the temperature gradient within the snowpack is sufficient. (Hoar crystals are formed by the sublimation of water vapour into ice, often building on an existing crystal. The water vapour comes from the air spaces between grains in the case of depth hoar, and from the air just above the snow surface in the case of surface hoar). Cold air over a thin snowpack is one condition that can create a temperature gradient capable of forming weak layers of depth hoar. Although layers of depth hoar can form at any level of the snowpack, the advanced forms are most commonly found near the bottom of the snowpack.
- Surface hoar, also called 'hoar frost', forms on cold, relatively clear nights. The crystals, some feather shaped and often larger than crystals that fall as snow, are very slow to change once buried in the snowpack. Typically, layers above and below the surface hoar will gain strength over a period of days or weeks, leaving the buried surface hoar as a relatively weak layer.

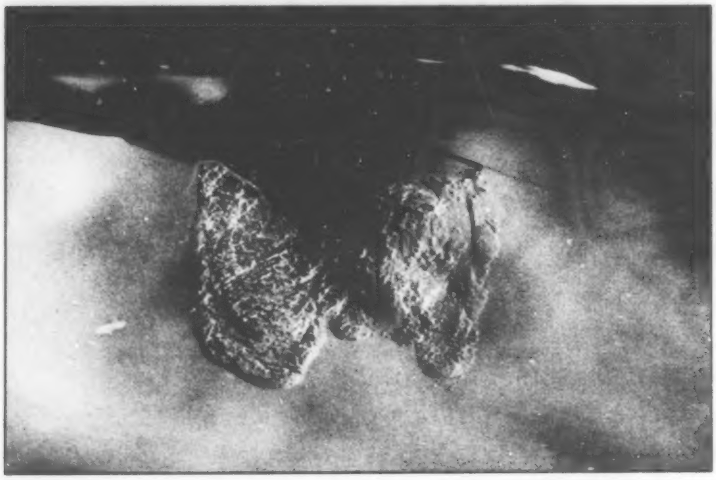
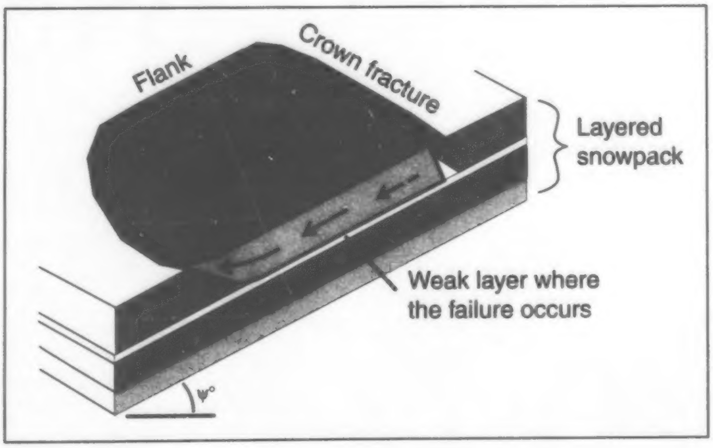


Figure 5.

A slab snow avalanche. Photograph by B. Jamieson. GSC 2000-103

Figure 6.

Slab avalanche nomenclature. The failures that release a slab avalanche start in the weak layer below the slab.



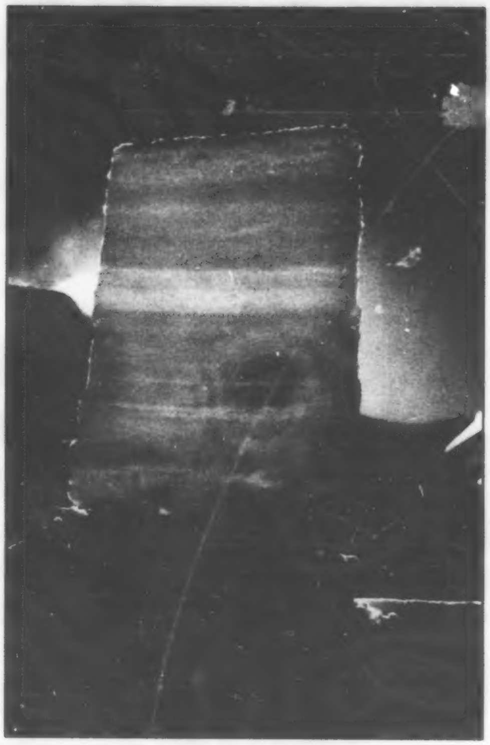


Figure 7. A translucent snowpack profile showing layers of varied thickness. Weak layers are usually less dense and more transparent than adjacent, stronger layers. Photograph by B. Jamieson. GSC 2000-104

Triggering

Most snow avalanches start naturally during or soon after snow storms. Others start when solar radiation, daytime heating, or relatively warm air masses warm the surface layers of the snowpack. On a slope, gravity causes the snowpack to creep downslope, with the top of the snowpack moving faster than the bottom. If the weak layer is deformed (sheared) fast enough, failure occurs in the weak layer, which produces the fractures in the perimeter of the overlying slab that are necessary to release the avalanche.

In at least 90% of recreational accidents, the victim or a member of his/her party triggers the avalanche. Field studies indicate that slab avalanches are triggered when the shear stress in the weak layer produced by the combined weight of the slab and the person exceeds the shear strength of the weak layer.

Some avalanches are triggered intentionally, usually by explosives used as part of avalanche control programs, but in places by skiers attempting to release unstable snow on short slopes in a controlled manner.

SNOW-AVALANCHE MOTION

Flow characteristics

Initially, slab-avalanche motion consists of blocks of snow gliding down the slope. Often within 10 to 50 m, the blocks begin to tumble and break up. At speeds above approximately 35 km/h, dry avalanches commonly generate a powder cloud (density 3–15 kg/m^3) that billows above the denser, flowing component (50–150 kg/m^3). Large dry slides can reach speeds of 50 to 200 km/h. Wet slides are denser (300–400 kg/m^3) and slower (20–100 km/h). In many cases, avalanches are so fast that they leave little time for a person downslope to get out of the way.

Victims who are engulfed in an avalanche should try to swim or struggle towards the surface where the forces are less and the odds of burial are reduced, or towards the side where the avalanche speed is less. Some avalanche victims who grab trees or rocks are left behind as the avalanche continues to travel downslope and thus avoid burial. Although the few people who have survived large avalanches report being helpless, there are many reports of such actions being effective in small- to medium-sized avalanches. When the avalanche slows, victims should struggle or thrust towards the surface. As the avalanche stops, victims should reach towards the surface and try to make an air space in front of their faces, possibly using one arm for each action.

AVALANCHE TERRAIN

A snow-avalanche path consists of a start zone, a track, and a runout zone where the avalanche decelerates and the snow is deposited. In some cases, these components of a path are easily identified (Fig. 8). Commonly on short slopes, the start zone, track, and runout are difficult to distinguish (Fig. 9).

Slope angle

The slope angle is the most important terrain factor influencing avalanche formation (Fig. 10). Slopes over 60° usually release numerous, small, loose snow avalanches, commonly called 'sluffs'. Although some sluffs are not large enough to bury a person, they can cause climbers to fall or can trigger larger avalanches on less steep slopes below.

Some large avalanches are released from slopes between 45° and 60°, but most occur on slopes between 30° and 45°. This is also the range of slope angles popular with skiers, snowboarders, and snowmobilers, hence the problem of recreationists triggering snow avalanches.

Less than 5% of dry slabs start on slopes under 30°. However, wet slides (i.e. slides in wet snow, which has liquid water between the grains of ice) can occur on slopes under 25°.

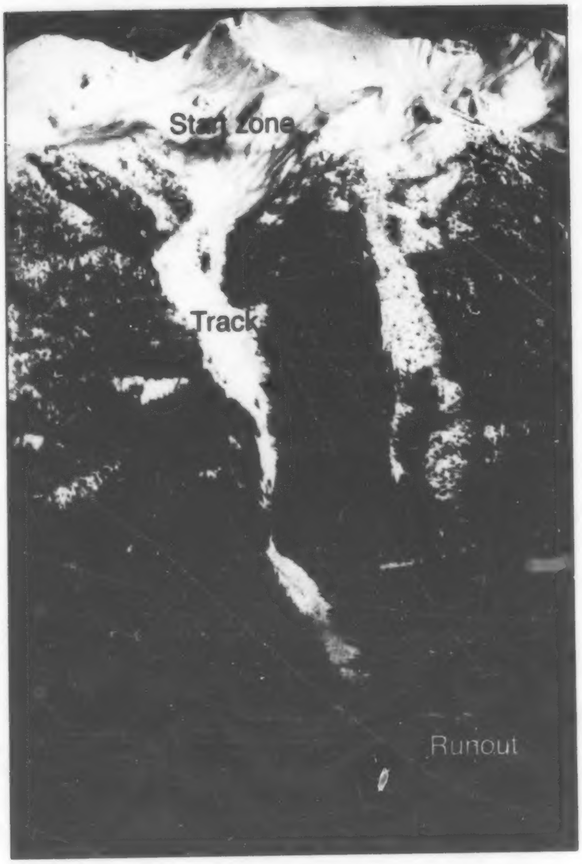

Start Zone
Track
Runout

Figure 8. A large avalanche path with an identifiable start zone above the treeline, track down the valley side and runout zone at the valley bottom. Photograph by B. Jamieson. GSC 2000-105

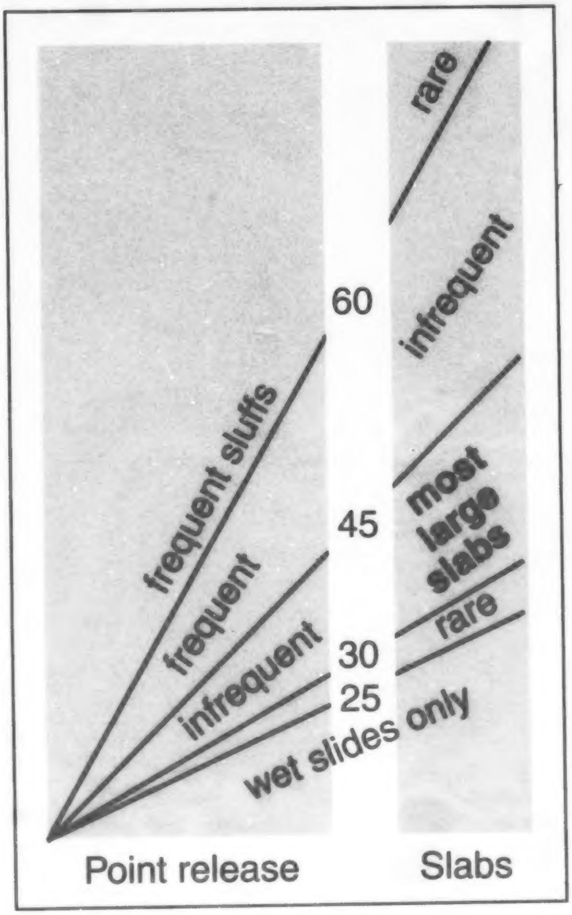

rare
infrequent
60
frequent sluffs
most large slabs
45
frequent
rare
infrequent
30
25
wet slides only
Point release
Slabs

***Figure 10.** Relationship between slope angle of the start zone and the character and frequency of avalanches. Steep slopes usually release small avalanches frequently. Most large, dry, slab avalanches start on slopes between 30° and 40°.*

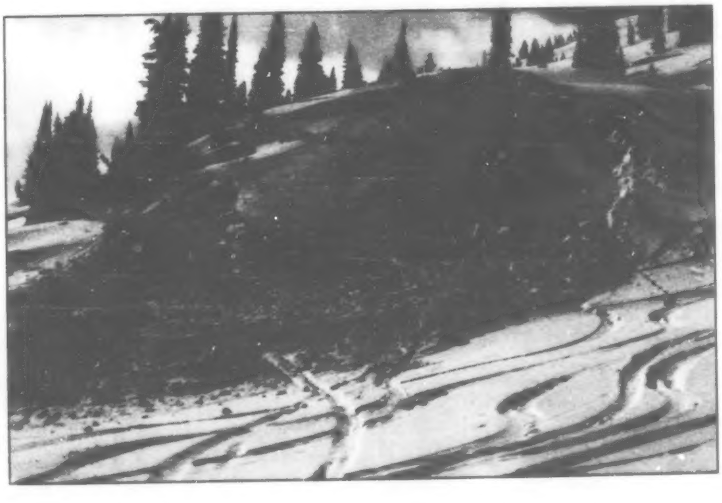


Figure 9.

A small avalanche slope with an indistinct start zone, track, and runout area. Many recreationists are caught in relatively small avalanche slopes. Photograph by B. Jamieson. GSC 2000-106

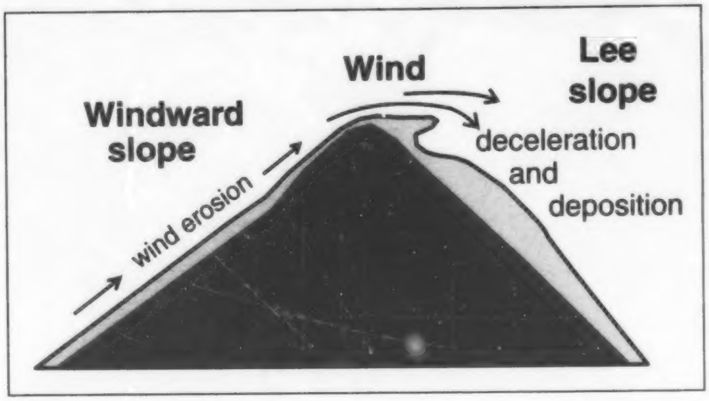


Figure 11.

Side view of a mountain ridge showing the usual effect of wind on snow distribution. Wind usually removes snow from windward slopes and deposits snow as slabs on lee slopes.

Table 1. Regional avalanche activity based on snow depth and terrain.

Colour	Terrain and average maximum snow depth (AMSD)	Expected avalanche activity
Red	mountainous and AMSD>100 cm	**High.** Numerous avalanche paths, many with avalanches most winters
Yellow	mountainous and 50<AMSD<100 cm, or many steep hills and AMSD>100	**Moderate.** Reduced density of avalanche paths, some with avalanches most winters
Green	many steep hills and 50<AMSD<100 cm	**Low.** Isolated avalanche slopes. Avalanches uncommon, mostly in extreme winters
White	mainly gentle terrain or AMSD<50 cm	**Very low.** Very isolated avalanche slopes. Avalanches rare

Orientation with respect to wind

The orientation of a slope with respect to wind strongly affects avalanche formation. Snow is picked up by wind on upwind slopes and some snow is deposited on the downwind or leeward slopes (Fig. 11). Wind deposits on the leeward slopes usually consist of stiff layers over weaker layers, reflecting the variable wind speed at the time of deposition. This creates ideal conditions for slab avalanches.

Other terrain factors

A variety of other terrain factors also influences avalanche formation. Sun-exposed slopes are commonly more prone to avalanche during sunny, warm conditions, whereas shady slopes are commonly less stable during cold, winter-like conditions. Rough ground such as talus fields requires a deeper snowpack before avalanching compared to slope surfaces of grass or smooth rock. Higher elevations commonly have a higher avalanche hazard because they have higher snowfall, more wind, and less vegetation than lower elevations. Wind slabs can accumulate on at least one side of gullies and bowls, which also funnel avalanches, increasing the destructive force and making escape difficult. These and other terrain factors are described in many books including Daffern (1992), McClung and Schaerer (1993), and Jamieson (1997).

AVALANCHE-PRONE AREAS IN CANADA

The most basic requirements for a snow avalanche are sufficient snow on the ground and a slope that is steep enough for an avalanche to slide. Generally, at least 50 cm of snow are required, and areas with deeper snow will usually avalanche more often.

A map depicting regional snow-avalanche activity in Canada is shown in Figure 12 (in pocket). The four levels of avalanche activity are based on terrain steepness and maximum annual snow depth as summarized in Table 1. The map shows generally that the areas of highest activity are in the mountain ranges in Western Canada and in northern Labrador. Most areas with moderate avalanche activity border areas of high activity in Western Canada, but some are also found in Labrador, in the Gaspé region of Quebec, along the north shore of the St. Lawrence River, and along the eastern margin of the Arctic Islands. Only low or very low activity is expected for large areas of the Prairies and for arctic, central, and Eastern Canada.

This map lacks the detail necessary for site-specific assessments of snow-avalanche activity or hazard as it does not take into account isolated steep areas, isolated areas of heavy snowfall or strong winds, extreme winters, unusual storms, or slush flows on less steep terrain. For example, the method used to delineate avalanche-activity areas underestimates activity on the Newfoundland coast, where windblown snow creates local avalanche areas.

Distribution of avalanche accidents

The locations of fatal avalanche accidents in Canada based on published sources (prior to June 1997) are listed in Appendix A and shown in Figure 12. Not surprisingly, most of these accidents occur in areas of high avalanche activity, particularly in Western Canada. Fewer occur in areas of moderate or low activity. Very few occur in remote areas such as northern Labrador where few people live and travel in avalanche terrain. Of course, fatal accidents can occur wherever a person is on a slope steep enough to avalanche given the right snow conditions. One example from the zone of very low activity occurred in 1978 when a man was killed by an avalanche in a gravel pit near Chelmsford, just north of Sudbury, Ontario.

Appendix A lists 216 snow-avalanche accidents involving 456 deaths. An additional 116 deaths occurred between 1885 and 1911 during the construction and early operation of the railway through Rogers Pass, British Columbia. However, the records giving the dates and locations of these additional accidents have been destroyed. In total, there are 572 reported avalanche fatalities in Canada. However, many fatal avalanche accidents, especially those prior to 1955, have not been reported. Currently, the Canadian Avalanche Centre in Revelstoke, British Columbia, collects reports of avalanches in which people were involved or property was damaged.

EXAMPLES OF AVALANCHE INCIDENTS

A recreational incident

On Sunday, 11 February 1990, a party of five skiers started up the Healy Creek trail near Banff, Alberta. A female skier lagged behind and then turned back during heavy snowfall and warm temperatures. She returned to the parking lot and waited for her friends who did not return. Later in the day, a group of skiers descending the trail crossed a huge avalanche deposit that had not been there earlier in the day. When they learned of the four missing skiers, they quickly searched the deposit and then returned to the trailhead where they reported the missing skiers.

The trail crosses the runout of two large avalanche paths. Although slides at this location usually do not reach the trail, avalanche conditions on this day were extreme. As temperatures warmed during the three-day snow storm, 76 cm of snow fell at Sunshine Village, 3 km away. The public bulletin released one day before the accident reported the avalanche hazard as *high*, but added "Snowfalls of up to 40 cm have fallen along the divide in the last 24 hours. Should heavy snowfalls forecast for Sunday materialize, the hazard will reach *extreme* in most areas of the park."

A search organized by Parks Canada wardens and involving many other agencies began that day and lasted for the next three days. The search party grew to include over 50 people, 2 helicopters, and several search dogs. During the search, the temperature dropped below -30°C and two searchers froze their toes. The dogs had difficulty locating the victims, probably due to the strong smell of coniferous trees smashed by the avalanche. The deceased victims were found on 13 and 14 February, 10 to 20 m into an area of mature timber beside the obvious avalanche runout zone.

At the time of the accident, the victims were having lunch in an area of mature forest that is normally considered safe. However, an unusually large avalanche of a type with a return period of 30 to 50 years occurred and extended laterally into the mature forest beside the obvious avalanche runout zone. The rapidly flowing snow pushed the victims 8 to 18 m through the forest and buried them 1.5 to 1.8 m below the surface. They were not wearing avalanche transceivers, homing devices that would have aided searchers in locating them.

A residential incident

Telegraph Creek is a small town on the Stikine River near the coast of northwestern British Columbia. Commonly in winter, more rain falls than snow. A mix of rain and wet snow fell in the town between 24 and 28 January 1989. Worried about the accumulation of wet snow on the slope above the eastern part of town, most residents abandoned their houses. One 80-year-old woman, however, declined to leave her house.

On the afternoon of 28 January, two avalanches swept down the slope, damaging three houses (Fig. 13, 14) and destroying the elderly woman's house. She was found alive in a chair in the remains of her house, but died soon after at a neighbour's home.

A railway incident

The Canadian Pacific Railway line through Rogers Pass, British Columbia, opened for regular traffic in 1886. During construction of the line and its operation prior to 1910, snow avalanches killed approximately 150 people. In terms of loss of life, the worst single accident occurred in 1910.

At 18:30 h on 4 March 1910, a large avalanche from Mount Cheops deposited about 7 m of dense snow on the tracks in the pass. A 64-person crew and a rotary snowplough attached to a locomotive were called in to clear the deposit. In spite of heavy snowfall, the work was considered safe because the Mount Cheops avalanche path had recently released an avalanche and there was mostly forest on the other side of the pass. The crew was cutting a 7 m deep trench through the dense snow deposit burying the tracks, when a large avalanche descended the slopes opposite Mount Cheops, smashing trees.

After reporting on progress at the watchman's shack at 23:30 h, the foreman noticed that the lights from the portable lanterns and plough could not be seen. Following a call for help, the foreman located the fireman of the locomotive. He had been sucked out the window of the locomotive and thrown 30 m onto an unused snow shed where he lay with

F2

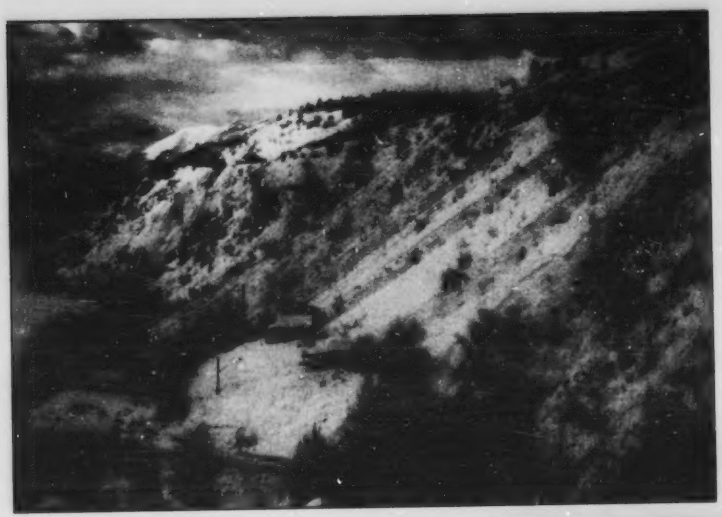


Figure 13.

Eastern part of Telegraph Creek, British Columbia, after the snow avalanche of January 1989. Note the house pushed off its foundation (arrow). Photograph courtesy of J. Bennetto

Figure 14.

Debris of a demolished house at Telegraph Creek, British Columbia. An 80-year-old woman was fatally injured in this house. Photograph courtesy of J. Bennetto

serious internal injuries, a badly broken leg, and dislocated shoulder. The foreman dragged him to a rail car and called to Revelstoke for help.

Two hundred people from Revelstoke, including nurses and doctors, travelled to Rogers Pass on a relief train, arriving at daylight the next morning. Shovelling through the dense deposit, they found most of the victims in the 7 m deep trench that had been filled by the second slide (Fig. 15). Sixty-two people died. In spite of his injuries, the fireman survived.

In response to this and previous accidents along the railway through Rogers Pass, the 8 km long Connaught Tunnel was constructed through Mount MacDonald to avoid some of the major avalanche paths.

Property damage

The Ruth Vermont gold mine operated intermittently from 1926. It is located along Vermont Creek in the Purcell Mountains, south of Golden, British Columbia. Much of the Vermont Creek valley bottom is exposed to avalanches from both sides of the valley. The mill and camp were grouped against a patch of trees on the south side of the creek. A larger and more exposed group of buildings was constructed in 1965. Avalanches in the winters of 1968 and 1970 each destroyed one of these buildings.

Beginning on 12 January 1974, 140 cm of snow fell at the mine over a four-day period. The wind was strong and temperatures approached 0°C. Early in the morning of 15 January, one avalanche from each side of the valley reached the mine, damaging the office building, mill, metals laboratory, oil shed, fuel tank, and a van. Towards midnight on the same day, one unoccupied bunkhouse was hit by another avalanche and a second building was partly destroyed. On the evening of 16 January, the powerhouse was hit by yet another avalanche. Without light or power, the miners retreated into the mine shaft. When they emerged in the morning, they found six buildings destroyed, six heavily damaged, and extensive equipment damage. No one was injured, but the cost of damage was estimated at $600 000.

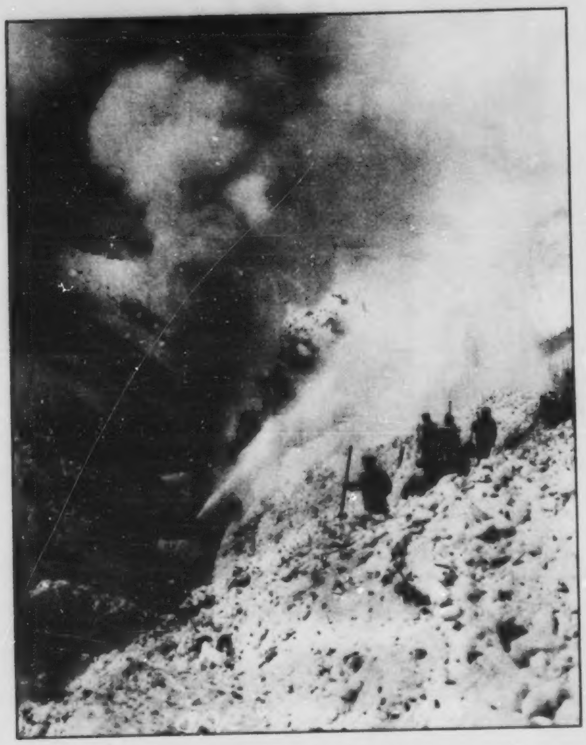


Figure 15. Crew clearing the avalanche deposit from the railway through Rogers Pass following the snow avalanche on 4 March 1910. Sixty-two people died in this avalanche. Photograph courtesy of Whyte Museum of the Canadian Rockies V263/NA71-1658 taken by Byron Harmon.

One of the remarkable aspects of the Ruth Vermont mine is the lack of serious avalanche accidents on the road, which is exposed to numerous avalanche paths. The mine operated for many years in winter without a reported fatality. Perhaps the mine manager or someone on staff doubled as an avalanche forecaster, interrupting hauling and crew changes during hazardous periods.

Damage to forests

On 14 March 1996, a wet slab avalanche began in a cutblock in Nagle Creek north of Revelstoke, British Columbia. The avalanche ran about 450 m down the cutblock, the slope of which averaged 37°. Small gullies split the avalanche into six separate avalanches that continued about 350 m through forest, destroying mature trees and depositing broken trees and topsoil in a lower, older clearcut (Fig. 16). The value of the damaged timber was approximately $400 000 (D. McClung, pers. comm., 1997). The avalanche caused substantial damage to eight-year-old saplings in the upper cutblock and damage to fifteen-year-old trees in the lower cutblock.

In order to prevent further destructive avalanches from starting in the upper cutblock, replanting in three strips starting at the top of the cutblock was recommended. This and other avalanche-prone cutblocks are being studied to develop harvesting strategies that minimize the risk of destructive avalanches.

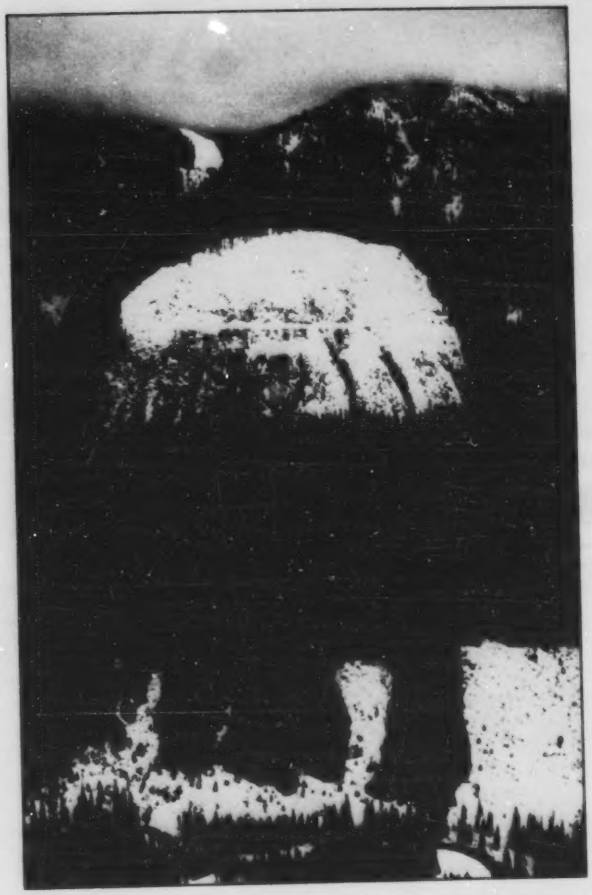


Figure 16. Cutblock avalanche in Nagle Creek north of Revelstoke, British Columbia. The photograph shows six swaths through mature timber between the upper and lower cutblocks. Photograph courtesy of Jim Bay.

HAZARD MITIGATION FOR PERMANENT STRUCTURES AND SKI AREAS

The following are methods used to mitigate avalanche hazards for permanent structures: location planning; structures in the start zone that prevent large avalanches from starting; structures in the track or runout zone that slow or deflect avalanches; reinforcement of the exposed structure; and forecasting and control programs (including temporary closures).

The primary protection for all permanent structures including transportation corridors, buildings, lift towers, and power lines is location planning. If a significant hazard remains after careful location of the structure, then other measures are necessary. For transportation corridors, secondary protection often involves a forecasting and control program that includes temporary closures. Deflecting structures and/or reinforcement often provide secondary protection for power lines and ski-lift towers.

Avalanche hazards at designated runs in a ski area are mitigated primarily by forecasting and control programs.

Location planning and zoning

Snow-avalanche hazards to structures are determined on the basis of estimates of avalanche frequency and magnitude at various locations within the avalanche path. This allows planners to place the structure or corridor in an area where the hazard is low, or to require implementation of other mitigative measures. Frequency is described in terms of the *return interval,* which is the estimated average number of years between avalanches of a certain size or extent.

For municipal land use, location planning involves identifying one or more hazard zones within the avalanche runout area. In Canada, commonly only one zone is identified by the 'avalanche hazard line'. Within this zone, destructive avalanches are expected with a return interval determined by the jurisdiction, e.g. 300 years. Currently, no consistent policies for avalanche hazard zoning exist in Canada.

In Switzerland and some jurisdictions in the United States, new residential buildings are usually not allowed in the 'red zone' where avalanches causing impact pressures of 30 kPa or more are expected with return intervals of less than 300 years, or any avalanches are expected with return intervals of less than 30 years. Some jurisdictions have a blue zone that extends beyond the red zone and typically includes areas with return intervals of between 30 and 300 years and impact pressures of less than 30 kPa. In this blue zone, the land manager or legislation may require that walls of a residence that face the avalanche path be reinforced or include a deflecting wedge. Beyond this zone, normal construction is allowed.

Start-zone structures

Structures installed in the start zone can support the snowpack sufficiently to prevent large avalanches. These structures usually resemble inclined fences or nets (Fig. 17). Although this might seem to be the ultimate solution, the structures are unsightly and so expensive that they are usually only used to protect permanent villages or busy roads. Less expensive wooden structures can be built to allow the regrowth of forest in the avalanche path. Experience in the design and maintenance of start-zone structures is considerable in the European Alps. However, in Canada, such structures are not common. Wooden 'fences' have been built in two small start zones east of Rogers Pass, British Columbia.

Where wind-deposited snow is the primary factor in avalanche formation, wind-modifying structures can be built that reduce the formation of wind slabs in the start zone. Some of these structures resemble robust versions of snow fences used in places along highways. These snow-collection fences create drifts on the gentle terrain above the avalanche start zone and reduce the amount of wind-blown snow that is deposited in the start zone. Other structures, which resemble sloping roofs, cause the wind to deposit the snow below the usual start zone.

Structures in the track and runout zone

A wide variety of structures are used to slow or deflect avalanches in the track or runout zone. Some of the more common structures are deflecting berms, splitting wedges, avalanche sheds, and mounds. Berms of rock or soil can deflect avalanches laterally away from buildings or parallel to the highway. Splitting wedges of soil or concrete constructed on the exposed side of a structure deflect the avalanche to both sides. Avalanche sheds or galleries can deflect avalanches over roads or railways. Mounds in the runout zone, usually at least 5 m high, can slow avalanches so they do not run as far. However, all such defense structure have their limitations. Occasionally, large avalanches will spill over the ends of avalanche sheds (Fig. 18) or the top of deflecting berms.

Figure 17.

Support structures in an avalanche start zone at Davos, Switzerland. Such structures can prevent large avalanches from starting, but because of their cost, are usually used to only to protect villages and busy transportation routes. These structures are rare in Canada. Photograph by B. Jamieson. GSC 2000-107

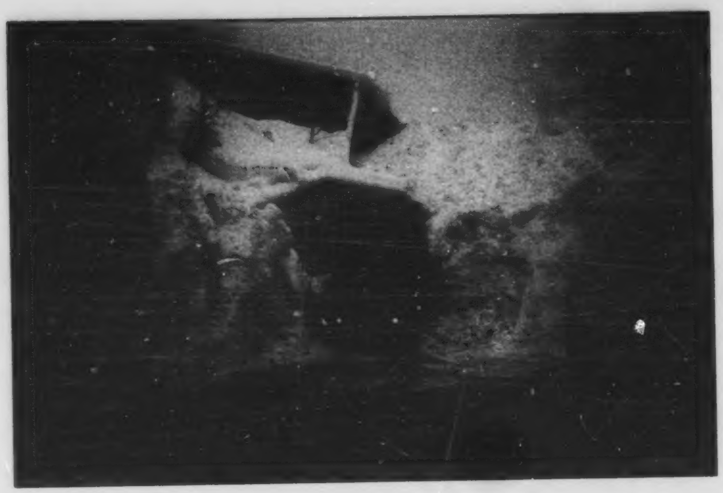


Figure 18.

The Lanark avalanche shed west of Rogers Pass, British Columbia, during snow removal after a large avalanche. Much of the avalanche shed and a second lane for traffic are buried under the deposit. During the design of this and other sheds in Rogers Pass, a spillover at the ends of the structure occurring about once every 10 years was considered acceptable in order to moderate the length and thus the cost of the shed. Photograph courtesy of Bruce Allen.

Mounds may be partly buried by deposits early in the winter, which allows subsequent avalanches to run largely unobstructed over them.

Control using explosives

Control programs for ski areas, highways, and railways often involve the use of explosives to test for unstable snow or to release unstable snow during temporary closures of specific areas. Charges, typically containing approximately 1 kg of explosive, can be projected from compressed-air canons or fired by military artillery (Fig. 19). Alternatively, charges with two-minute or longer fuses are dropped into start zones from ridges or from helicopters. Approximately 30 000 charges are used each year for avalanche control in Canada, about half of which are used in ski areas.

Although explosions in start zones *usually* release unstable snow, they are not a guarantee that the slope will not release naturally or be triggered by a person soon afterwards. The results of the use of explosives are only one factor considered by snow-avalanche forecasters.

Forecasting

Some snow-avalanche forecasters must make decisions concerning the opening and closing of portions of transportation corridors or ski areas. Others in parks or information centres provide information and danger ratings for the backcountry. Although forecasters for transportation corridors are most interested in natural avalanches that might reach a highway or railway line, forecasters for ski areas and the backcountry focus on avalanches, especially those triggered by people, that are large enough to injure or bury recreationists.

Avalanche forecasting involves an assessment of the likelihood and size of snow avalanches under existing or future conditions. Forecasts are based on the following categories of information: 1) observation of snow avalanches, including the observation of few or no avalanches; 2) stability and strength tests, including the explosive test, the shovel test

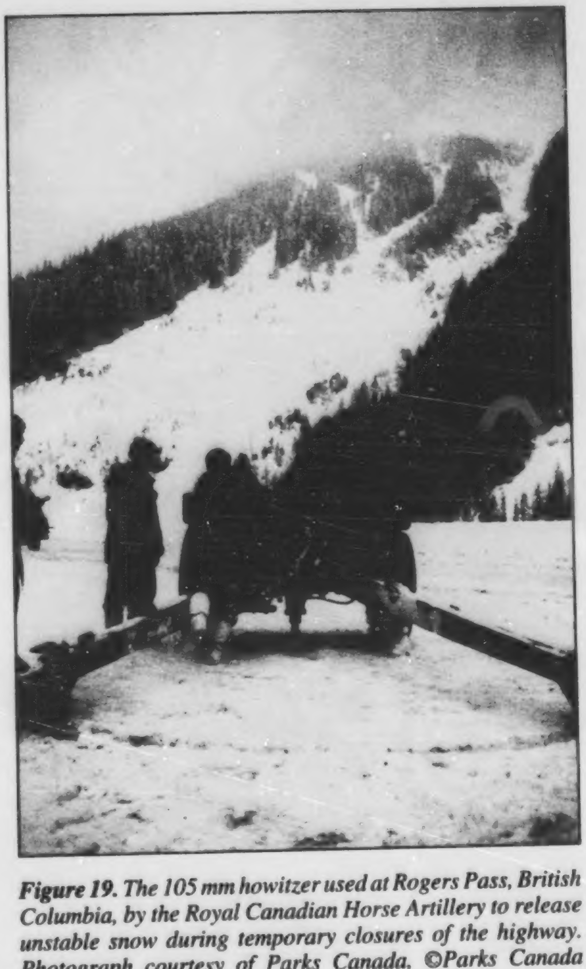


***Figure 19.** The 105 mm howitzer used at Rogers Pass, British Columbia, by the Royal Canadian Horse Artillery to release unstable snow during temporary closures of the highway. Photograph courtesy of Parks Canada, ©Parks Canada #615-288-O-577*

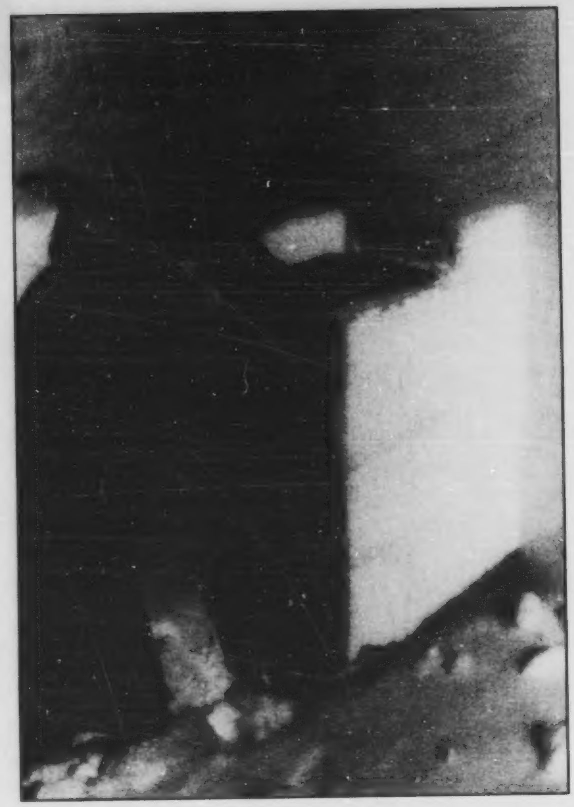


Figure 20. A compression test used to locate weak layers in the snowpack and provide an index of stability. Note the weak layer that has fractured halfway down the column (darker band). Photograph by B. Jamieson. GSC 2000-108

(a traditional test in which a shovel is used to pull on the upper part of a column of undisturbed snow, roughly 0.3 m square; fractures produced along weak snowpack layers identify potential failure planes for slab avalanches), the compression test (Fig. 20), or the rutschblock test (a snow-stability test in which a squat column of undisturbed snowpack, 2 m across the slope and 1.5 m along the side, is loaded by a skier stepping onto the column, pushing down with the legs, then jumping on the column; the score is the loading stage, from 1 to 7, that causes the upper part of the column to release a block); 3) snowpack observations, including profiles that identify the presence, grain structure, and properties of weak layers; 4) weather observations, especially observations of precipitation, wind and temperature.

The observation of recent avalanches provides the most direct evidence of the likelihood of avalanches occurring; weather observations provide the least direct evidence. Stability tests and snowpack observations fall in between.

Forecasters are interested in observations that indicate unstable or stable snowpack conditions. For example, observations of recent avalanches, weak layers in the snowpack, heavy snowfall, and warming would indicate that avalanches are likely. Alternatively, if no or few avalanches have been observed, if stability tests suggest stability, if no well defined weak layers are apparent in the snowpack, and if the weather is cold and clear, a forecaster will probably forecast a low danger of avalanches and allow the highway or ski area to remain open. However, these two examples are simplistic. In many cases, forecasters must make decisions on the basis of some factors that suggest stability and other factors that indicate instability. They consider the various factors, weighting some more than others, and assess the uncertainty associated with the observations and the risk associated with the possible forecasts. The decision, however, is ultimately based on their experience and training. Forecasts are routinely revised as more information is gathered.

Recently, numerical and rule-based computer systems have been developed to assist forecasters. The best known numerical system, called 'Nearest Neighbours', compares current measurements of precipitation, wind, temperature, and other variables with values in the database for that area that span many years. If avalanches were reported on most past days with conditions similar to current values, then avalanches are predicted to be likely. Alternatively, if no avalanches were recorded on most of the days with similar conditions, then avalanches are considered to be unlikely.

Rule-based or expert systems incorporate the knowledge of snow avalanche experts. Rules are constructed about the effect of snow stratigraphy patterns, precipitation, wind, temperature, field tests, and other factors relating to snow stability and/or avalanche danger. The following are two examples of simplified rules:

- increase avalanche danger by one level if wind speed exceeds 25 km/h;
- decrease danger one level if temperature has dropped below 0°C.

No previous or historical data are required to develop such rules. The system of rules is then tested with current data and revised if necessary.

AVALANCHE SAFETY FOR BACKCOUNTRY RECREATION

When travelling in the mountains in winter, recreationists commonly use the following questions to assess and minimize the local snow avalanche danger: 1) Is the slope to be travelled prone to avalanches? 2) Is the snowpack unstable? 3) What are the consequences of being caught by an avalanche on this terrain?

The first question is answered using terrain factors such as slope angle and orientation to wind that were outlined earlier. The second question concerning snow stability is answered using public bulletins, avalanche activity, stability tests, snowpack observations, and weather observations. The third question requires the consideration of a combination of snowpack and terrain factors: Is a large avalanche likely? Could I be swept over a cliff, or into a crevasse, trees, or boulders?

Unfortunately, certain inappropriate human factors too often enter into decisions. Some people focus on a mountain peak, attractive slope, or other destination and may continue in spite of strong evidence of avalanche danger. Groups may spread out, thus excluding slower, but capable, people at the back from decisions about the route or avalanche danger. At the end of the day when people are tired, careful decision making may be forgotten in the rush to get back. Also, many avalanche accidents happen during fair weather as such weather draws recreationists towards open slopes, high mountain passes, and beautiful summits.

Winter recreationists, particularly less experienced ones, can reduce their risk by travelling with more experienced people or hiring a guide.

Appendix B presents a snow-avalanche danger scale provided by the Canadian Avalanche Association.

AVALANCHE SURVIVAL AND RESCUE

The motion of a snow avalanche fatally injures approximately 25% of avalanche victims, and after 30 minutes of burial the survival rate drops to 50%. Consequently, to have a 50% chance of finding a buried victim alive, his or her head and chest must be dug out within 30 minutes. Only about 15% of victims survive being buried two hours. In most backcountry situations, few victims will survive long enough for surviving members of the group to get help. The victim's best chance for survival depends on an effective search by the other members of the group. Reports of live recoveries by organized searchers called to the accident site are rare. Victims trapped in partly collapsed buildings have, on rare occasions, survived for several days.

The initial rescue response depends on the available resources and on whether the accident occurred in the backcountry, on a road, or in a residential or industrial area.

If the victims are wearing avalanche transceivers, searchers can switch their transceivers to receive mode and locate most victims within 5 to 15 minutes. This does not include the time needed to dig the victim out, which can take as long or longer. Transceivers work! Most winters in Canada, a few buried victims are reported found using transceivers and recovered alive. Search time for avalanche dogs is comparable to the time needed to find a person with a transceiver, although a trained dog rarely arrives at an accident site in time. (There are two reports of buried victims being found alive by an avalanche dog in the United States, but none in Canada.) If the victim(s) and the survivors are not using transceivers and a dog is not at the accident site, then survivors should probe for the victim(s) using a proven probing pattern. However, five probers are approximately 100 times slower than one searcher with a transceiver. Sectional probes work best, although people have been found by probing with branches, skis, or ski poles.

The odds of survival decrease considerably with increasing depth of burial. Only 5 to 10% of avalanche victims survive being buried in more than 1.5 m of snow. The grim odds for people caught in avalanches emphasize the importance of the mitigative measures described above. For recreationists, information on current conditions, avalanche education, and a thoughtful decision-making process are essential.

ACKNOWLEDGMENTS

I am grateful to Peter Schaerer, Jim Bay, Mike Boissonneault, Alan Dennis, Christoph Dietzfelbinger, George Evanoff, Torsten Geldsetzer, David Liverman, David McClung, and Chris Stethem for their comments on certain accidents or hazards, and to Ron Perla for translating some Russian text. Eric Tull and the staff at the Map Library of The University of Calgary were very helpful during my search for the locations of the accidents. Information sources for this chapter include those listed under 'Suggested reading'. My thanks to Greg Brooks and Larry Dyke of the Geological Survey of Canada who reviewed this contribution.

REFERENCES

Batterson, M., Liverman, D., and Taylor, D.
1995: The assessment of geological hazards and disasters in Newfoundland; Newfoundland Department of Natural Resources, Geological Survey, Report 95-1, p. 55–75.

Daffern, T.
1993: Avalanche Safety for Skiers and Climbers (second edition); Rocky Mountain Books, Calgary, Alberta, 192 p.

Jamieson, B.
1997: Backcountry Avalanche Awareness (sixth edition); Canadian Avalanche Association, Revelstoke, British Columbia, 47 p.

Jamieson, J.B and Geldsetzer, T.
1996: Avalanche Accidents in Canada. Volume 4: 1984–1996; Canadian Avalanche Association, Revelstoke, British Columbia, 202 p.

Manners, E.
1997: Avalanche Involvements 1996–97; Avalanche News 52, Canadian Avalanche Association, Revelstoke, British Columbia, p. 22–24.

McClung, D.M. and Schaerer, P.A.
1993: The Avalanche Handbook; The Mountaineers, Seattle, Washington, 271 p.

McFarlane, R.C.
1985: Snow avalanche impacts and management in Canada; Ph.D. thesis, University of Waterloo, Waterloo, Ontario, 272 p.

Schaerer, P.A.
1987: Avalanche accidents in Canada III, a selection of case histories, 1978–1984; National Research Council of Canada, Institute for Research in Construction, Publication 27950, 138 p.

Stethem, C.J. and Schaerer, P.A.
1979: Avalanche accidents in Canada I, a selection of case histories of accidents, 1955 to 1976; National Research Council of Canada, Publication 17292, 114 p.
1980: Avalanche accidents in Canada II, a selection of case histories of accidents, 1943 to 1978; National Research Council of Canada, Publication 18525, 75 p.

SUGGESTED READING

Textbooks and reviews

Armstrong, B.R. and Williams, K.
1992: The Avalanche Book; Fulcrum Publishing, Golden, Colorado, 240 p.

Mellor, M.
1968: Avalanches; Cold regions science and engineering, Part III-A3; Snow Technology; United States Army Cold Regions Research and Engineering Laboratory, 224 p.

Perla, R.I.
1980: Avalanche release, motion and impact; *in* Dynamics of Snow and Ice Masses, (ed.) S.C. Colbeck; Academic Press, New York, p. 397–462.

Schaerer, P.A.
1981: Avalanches; *in* Handbook of Snow: Principles, Processes, Management and Use, (ed.) D.M. Gray and D.H. Male; Pergamon Press, p. 475–516.

Avalanche safety for recreationists

Fredston, J. and Fesler, D.
1994: Snow Sense: a Guide to Evaluating Snow Avalanche Hazard; Alaska Mountain Safety Center, Inc., Anchorage, Alaska, 116 p.

Fredston, J., Fesler, D., and Tremper, B.
1995: The human factor — lessons for avalanche education; Proceedings of the International Snow Science Workshop at Snowbird, International Snow Science Workshop 1994, Snowbird, Utah, p. 473–486.

Jamieson, B. and McDonald, J.
1999: Free Riding in Avalanche Terrain — A Snowboarder's Handbook; Canadian Avalanche Association, Revelstoke, British Columbia, 80 p.

Jamieson, B. and Svederus, D.
1998: Snowmobiling in Avalanche Terrain — Reducing the Risk; Canadian Avalanche Association, Revelstoke, British Columbia, 56 p.

Avalanche forecasting

Buser, O.
1989: Two years experience of operational avalanche forecasting using the nearest neighbours method; Annals of Glaciology, v. 13, p. 31–34.

Föhn, P.M.B.
1987: The rutschblock as a practical tool for slope stability evaluation; Avalanche Formation, Movement and Effects, International Association of Scientific Hydrology Publication 162 (Symposium at Davos, Switzerland, 1986), p. 223–228.

Jamieson, J.B. and Johnston, C.D.
1995: Interpreting rutschblocks in avalanche start zones; Avalanche News, v. 46, p. 2–4.

LaChapelle, E.R.
1980: The fundamental processes in conventional avalanche forecasting; Journal of Glaciology, v. 26, p. 75–84.

McClung, D.M. and Tweedy, J.
1994: Numerical avalanche prediction: Kootenay Pass, British Columbia, Canada; Journal of Glaciology, v. 40, p. 350–358.

Mears, A.I.
1992: Snow-avalanche hazard analysis for land-use planning and engineering; Colorado Geological Survey, Bulletin 49, 55 p.

Perla, R.I.
1970: On contributory factors in avalanche hazard evaluation; Canadian Geotechnical Journal, v. 7, p. 414–419.

Schweizer, J. and Föhn, P.M.B.
1996: Avalanche forecasting — an expert system approach; Journal of Glaciology, v. 42, p. 318–332.

Economic impacts

Morrall, J.
1995: Cost of traffic delays due to avalanche related closures on the Trans-Canada Highway west of Revelstoke; Snow Avalanche Programs, British Columbia Ministry of Transportation and Highways, 13 p.

Morrall, J.F. and Abdelwahab, W.M.
1992: Estimating traffic delays and the economic cost of recurrent road closures on rural highways; Logistics and Transportation Review, v. 29, p. 159–177.

Geography of avalanches

Fisheries and Environment Canada
1978: Hydrological Atlas of Canada — Atlas hydrologique du Canada; Ministry of Supply and Services, Ottawa.

Mlagkov, S.M. and Kanaev, L.A. (ed.)
1992: Geography of Avalanches; Moscow State University, Moscow, Russia, 332 p. (in Russian).

Schaerer, P.A.
1984: Estimates of the number of avalanches in western Canada; Avalanche News, v. 16, p. 3–6.

Woods, J.
1992: Snow War, an Illustrated History of Rogers Pass, Glacier National Park (third edition); Canadian Parks and Wilderness Society, Toronto, Ontario, 52 p.

Snow properties

Brun, E., Martin, E., Simon, V., Gendre, C., and Coleou, C.
1989: An energy and mass model of snow cover suitable for operational avalanche forecasting; Journal of Glaciology, v. 35, p. 333–342.

Canadian Avalanche Association
1995: Observation Guidelines and Recording Standards for Weather, Snowpack and Avalanches; Canadian Avalanche Association, P.O. Box 2759, Revelstoke, British Columbia, 98 p.

Colbeck, S.C.
1991: The layered character of snow covers; Reviews of Geophysics, v. 29, p. 81–96.

Colbeck, S., Akitaya, E., Armstrong, R., Gubler, H., Lafeuille, J., Lied, K., McClung, D., and Morris, E.
1990: International Classification for Seasonal Snow on the Ground; International Commission for Snow and Ice (International Association of Hydrological Sciences), World Data Center A for Glaciology, University of Colorado, Boulder, Colorado, 23 p.

LaChapelle, E.R.
1969: Field Guide to Snow Crystals; University of Washington Press, Seattle, Washington, 101 p.

Perla, R.I.
1977: Slab avalanche measurements; Canadian Geotechnical Journal, v. 14, p. 206–213.

Stethem, C. and Perla, R.
1980: Snow-slab studies at Whistler Mountain, British Columbia, Canada; Journal of Glaciology, v. 26, p. 85–91.

Incidents, survival, and rescue

Burtscher, M.
1994: Avalanche survival chances; Nature, v. 371, no. 6497, p. 482.

Föhn, P.M.B.
1987: The stability index and various triggering mechanisms; *in* Avalanche Formation, Movement and Effects, (ed.) B. Salm and H. Gubler; International Association of Hydrological Sciences, Publication no. 162, p. 195–211.

Jamieson, J.B. and Johnston, C.D.
1992: Snowpack characteristics associated with avalanche accidents; Canadian Geotechnical Journal, v. 29, p. 862–866.

Ommanney, C.S.L.
1978: Major Canadian avalanches 1970–76 and the UNESCO annual reports; *in* Avalanche Control, Forecasting and Safety, Proceedings of a Workshop in Banff, Alberta, November 1976, p.195–198.

Schaerer, P.A.
1988: Studies on survival in avalanches; Canadian Alpine Journal, v. 71, p. 55–56.

APPENDIX A

Fatal avalanche accidents in Canada (1863–June 1997)

Date (y/m/d)	Location	Province	Fatalities	Source
1863/03/12	Distress Cove, Placentia Bay	Newfoundland	2 hunters	Batterson et al. (1995)
1877/02/09	Betts Cove	Newfoundland	6 residents	Batterson et al. (1995)
1877/02/09	Betts Cove	Newfoundland	1 RCMP sergeant	D. Liverman (pers. comm., 1998)
1885/02/08	McDermot's Camp, Rogers Pass	British Columbia	3 workers	McFarlane (1985)
1885/02/25	Rogers Pass summit	British Columbia	6 workers	McFarlane (1985)
before 1889	unknown	British Columbia	1 miner	McFarlane (1985)
1895/01	Bluff Head Brooks	Newfoundland	4 miners	D. Liverman (pers. comm., 1998)
1898/02/27	road to Silverton mine	British Columbia	1 miner	McFarlane (1985)
1899/01/30	Rogers Pass avalanche shed	British Columbia	1workers	McFarlane (1985)
1899	Glacier CPR station	British Columbia	8 persons	McFarlane (1985)
1902/12/25	Molly Gibson mine near Slocan	British Columbia	9 miners	McFarlane (1985)
1904/01/25	near Revelstoke	British Columbia	1 workers	McFarlane (1985)
1904/01/30	50 km north of Squamish	British Columbia	1 trapper	McFarlane (1985)
1904/04/11	1 km east of Revelstoke	British Columbia	2 workers	McFarlane (1985)
1907/01/22	Devlin Group near Salmo	British Columbia	1 miner	McFarlane (1985)
1907/12/24	4 km east of Burton City	British Columbia	2 persons on road	McFarlane (1985)
1910/03/04	Avalanche Crest, Rogers Pass	British Columbia	62 workers	McFarlane (1985)
1911/12/23	Noble No. 5 mine near Sandon	British Columbia	1 miner	McFarlane (1985)
1912/03/11	Tilt Cove	Newfoundland	1 mine manager and 4 residents	Batterson et al. (1995)
1912/12/30	Coal Creek near Fernie	British Columbia	6 persons	McFarlane (1985)
1913/01/06	Noble No. 5 mine near Sandon	British Columbia	3 miners	McFarlane (1985)
1913/01/08	Dunedin mine near Sandon	British Columbia	1 miner	McFarlane (1985)
1913/01/22	Coal Creek near Fernie	British Columbia	1 person	McFarlane (1985)
1917/12/31	Shellbird Island, Humber River	Newfoundland	1 section foreman	Batterson et al. (1995)
1917/12/??	near Hazelton	British Columbia	2 persons	McFarlane (1985)
1921/02/07	Signal Hill, St. John's	Newfoundland	1 resident	Batterson et al. (1995)
1927	Mount Serra near Vancouver	British Columbia	1 climber	McFarlane (1985)
1927/02/24	Extenuate mine near Stewart	British Columbia	1 miner	McFarlane (1985)
1927/12/17	Porter Idaho mine near Marmot River	British Columbia	1 miner	McFarlane (1985)
1930/08/05	unknown		1 climber	McFarlane (1985)
1933/03/??	Duchesnay Pass, Yoho National Park	Alberta	2 skiers	McFarlane (1985)
1933/04/07	Fossil Mountain, Banff National Park	Alberta	1 skier	McFarlane (1985)
1935/03/01	Okak, Labrador	Newfoundland	2 residents	Batterson et al. (1995)
1935/03/04	Curling Road, Corner Brook	Newfoundland	3 residents	Batterson et al. (1995)
1937/03/??	Rogers Pass	British Columbia	1 worker	McFarlane (1985)
1943/02/11	McLean Point., Kwinitsa, Skeena River valley	British Columbia	3 workers	Stethem and Schaerer (1980)
1945/03/26	Mount Richardson, Banff National Park	Alberta	1 skier	McFarlane (1985)
1948/12/30	Upper Capilano near Vancouver	British Columbia	1 city employee	McFarlane (1985)
1949/02/22	Hoodoo Curve near Boston Bar	British Columbia	1 driver	Stethem and Schaerer (1980)
1950/01/30	British Columbia interior	British Columbia	2 workers	McFarlane (1985)
1950/02/13	Hell's Gate, Fraser River	British Columbia	1 worker	McFarlane (1985)
1950/02/25	near Flood	British Columbia	1 person on road	McFarlane (1985)
1955/03/27	Seymour Inlet, northwest of Vancouver	British Columbia	1 logger	Stethem and Schaerer (1980)
1955/07/11	Mount Temple, Banff National Park	Alberta	7 climbers	Stethem and Schaerer (1979)
1956/03/11	Marmot Peak, Jasper National Park	Alberta	1 skier	Stethem and Schaerer (1979)
1957/02/23	Woodbury Creek, Kootenay Lake	British Columbia	1 person on road	Stethem and Schaerer (1980)
1957/02/23	3 km south of Kaslo	British Columbia	1 person on road	McFarlane (1985)
1957/03/17	Richardson Ridge near Lake Louise	Alberta	1 skier	Stethem and Schaerer (1979)
1958/11/23	Bow Summit, Banff National Park	Alberta	2 skiers	Stethem and Schaerer (1979)
1959/02/16	The Battery, St. John's	Newfoundland	5 residents	Batterson et al. (1995)
1959/03/24	McGillivray Pass	British Columbia	2 skiers	Stethem and Schaerer (1980)
1959/04/07	Torbit mine, Alice Arm	British Columbia	1 worker	Stethem and Schaerer (1980)

Date (y/m/d)	Location	Province	Fatalities	Source
1960/07/30	Mount Waddington, Coast Mountains	British Columbia	4 climbers	McFarlane (1985)
1961/07/19	Saddle Peak, Mount Garibaldi	British Columbia	1 climber	Stethem and Schaerer (1980)
1962/01/21	Silver Star	British Columbia	1 skier	Stethem and Schaerer (1980)
1962/02/06	Griquet, northern peninsula	Newfoundland	1 tobogganer	D. Liverman (pers. comm., 1998)
1962/12/30	Mount Whaleback, Yoho National Park	British Columbia	1 skier	Stethem and Schaerer (1979)
1965/01/??	between Darcy and Lillooet	British Columbia	1 worker	McFarlane (1985)
1965/02/13	Ocean Falls	British Columbia	7 persons in buildings	Stethem and Schaerer (1980)
1965/02/18	Granduc mine, 30 km northwest of Stewart	British Columbia	26 workers	Stethem and Schaerer (1979)
1965/02/28	Mount Norquay	Alberta	1 skier	Stethem and Schaerer (1979)
1965/12/28	Kootenay Pass	British Columbia	1 person on road	Stethem and Schaerer (1979)
1966	Red Pass	British Columbia	2 persons on road	Stethem and Schaerer (1979)
1966/01/08	MacDonald West #2, Rogers Pass	British Columbia	2 workers	Stethem and Schaerer (1979)
1966/01/08	national parks	British Columbia	1 worker	McFarlane (1985)
1967/01/30	Ptarmigan Chutes, Mount Whitehorn	Alberta	1 skier	Stethem and Schaerer (1979)
1967/12/09	Parkers Ridge, Banff National Park	Alberta	1 skier	Stethem and Schaerer (1979)
1969/03/23	Mount Hector, Banff National Park	Alberta	1 skier	Stethem and Schaerer (1979)
1970/01/24	Westcastle near Pincher Creek	Alberta	1 skier	Stethem and Schaerer (1979)
1971/08/11	Mount St. Elias	Yukon Territory	4 climbers	Stethem and Schaerer (1980)
1971/12/23	Granite Mountain near Rossland	British Columbia	1 skier	Stethem and Schaerer (1979)
1971/12/23	30 km east of Fernie	British Columbia	3 loggers	McFarlane (1985)
1971/12/24	Flathead River valley	British Columbia	3 persons on road	Stethem and Schaerer (1979)
1972/02/05	Highland Creek, Scarborough	Ontario	2 tobogganers	Stethem and Schaerer (1979)
1972/02/19	Mount Edith Cavel, Jasper National Park	Alberta	3 climbers	Stethem and Schaerer (1979)
1972/03/05	Giant Mascot mine, 6 km from Hope	British Columbia	3 persons on road	Stethem and Schaerer (1979)
1972/04/08	Whistler Mountain	British Columbia	4 skiers	Stethem and Schaerer (1979)
1972/04/23	Apex Basin, Ymir Mountain, Nelson	British Columbia	1 skier	Stethem and Schaerer (1979)
1973/03/14	Nine Mile Mountain, northeast of Hazelton	British Columbia	1 snowmobiler	Stethem and Schaerer (1980)
1973/12/09	Bow Summit, Banff National Park	Alberta	1 skier	Stethem and Schaerer (1979)
1974/01/22	restaurant, 45 km west of Terrace	British Columbia	7 persons in building	Stethem and Schaerer (1979)
1974/02/07	Bonnet Plume Range	Yukon Territory	1 climber	McFarlane (1985)
1974/02/17	Mica Mountain, Cariboo Mountains	British Columbia	1 skier	Stethem and Schaerer (1979)
1974/03/30	Lookout Mountain, Banff National Park	Alberta	1 skier	Stethem and Schaerer (1979)
1974/08/29	Mount Weisshorn, St. Elias Range	Yukon Territory	2 climbers	McFarlane (1985)
1976/01/14	Cap Santé, comté Portneuf	Quebec	1 tobogganer	Stethem and Schaerer (1979)
1976/01/16	Kootenay Pass, west side	British Columbia	3 persons on road	Stethem and Schaerer (1979)
1976/03/23	Paradise Basin, Invermere	British Columbia	1 skier	Stethem and Schaerer (1979)
1976/12/12	Chancellor Peak, Yoho National Park	Alberta	3 climbers	Stethem and Schaerer (1980)
1976/12/28	5 km east of Tadoussac, comté Dubuc	Quebec	1 snowmobiler	Stethem and Schaerer (1980)
1977/02/15	Parkers Ridge, Jasper National Park	Alberta	1 skier	Stethem and Schaerer (1980)
1977/03/16	Diana Lake, Table Mountain Ridge	British Columbia	1 skier	Stethem and Schaerer (1980)
1977/03/17	Bugaboo Mountains south of Golden	British Columbia	3 skiers	Stethem and Schaerer (1980)
1977/03/19	Bow Peak, Banff National Park	Alberta	1 climber	Stethem and Schaerer (1980)
1977/03/27	Quartz Ridge	Alberta	1 skier	Stethem and Schaerer (1980)
1977/03/30	Flush Bowl near Lake Louise ski area	Alberta	1 skier	Stethem and Schaerer (1980)
1977/10/09	President's Glacier, Yoho National Park	British Columbia	1 climber	Stethem and Schaerer (1980)
1978/01/29	Chelmsford	Ontario	1 snowshoer	Schaerer (1987)
1978/01/29	Mount Bowlen near Moraine Lake	Alberta	1 climber	Schaerer (1987)
1978/01/29	Dennis Creek near Slocan Lake	British Columbia	1 skier	Schaerer (1987)
1979/02/14	Spillimacheen Range	British Columbia	7 skiers	Schaerer (1987)
1979/02/20	Back Bowl of Whistler Mountain	British Columbia	1 skier	Schaerer (1987)
1979/02/23	Mount McKenzie near Revelstoke	British Columbia	1 skier	Schaerer (1987)
1979/02/24	Stanley Basin, Kootenay National Park	British Columbia	1 skier	Schaerer (1987)

Appendix A (cont.)

Date (y/m/d)	Location	Province	Fatalities	Source
1979/02/24	Mount Kitchener, Jasper National Park	Alberta	1 climber	Schaerer (1987)
1979/02/28	Tangle Hill, Jasper National Park	Alberta	1 climber	Schaerer (1987)
1979/03/17	Windy Pass near Gold Bridge	British Columbia	1 snowmobiler	Schaerer (1987)
1979/08/20	Mount Bowlen near Moraine Lake	Alberta	1 climber	McFarlane (1985)
1980/03/15	Mount MacKenzie near Revelstoke	British Columbia	1 skier	Schaerer (1987)
1980/08/08	Slesse Mountain	British Columbia	1 climber	Schaerer (1987)
1980/08/20	Mount Athabasca, Jasper National Park	Alberta	1 climber	Schaerer (1987)
1981/02/18	Derickson Ridge near Revelstoke	British Columbia	1 skier	Schaerer (1987)
1981/02/21	Waterfall Creek, Yoho National Park	British Columbia	1 skier	Schaerer (1987)
1981/02/22	west of Borgeau Lake, Banff National Park	Alberta	1 skier	Schaerer (1987)
1981/02/22	Mount Thompson, Banff National Park	Alberta	2 skiers	Schaerer (1987)
1981/02/23	Purcell Mountains south of Golden	British Columbia	3 skiers	McFarlane (1985)
1981/04/02	Mount Stephen, Yoho National Park	British Columbia	1 climber	Schaerer (1987)
1982/01/12	70 km west of Terrace	British Columbia	1 worker	Schaerer (1987)
1982/01/18	Purcell Mountains south of Golden	British Columbia	2 skiers	Schaerer (1987)
1982/02/05	Cirrus Mountain, Banff National Park	Alberta	1 climber	Schaerer (1987)
1982/02/22	Marmot Peak, Jasper National Park	Alberta	1 skier	Schaerer (1987)
1982/06/11	Mount Logan	Yukon Territory	3 climbers	Schaerer (1987)
1982/08/19	Crescent Glacier, Bugaboo Spire	British Columbia	1 climber	Schaerer (1987)
1982/08/20	Mount Robson	British Columbia	2 climbers	Schaerer (1987)
1982/08/22	Mount Kitchener, Jasper National Park	Alberta	1 climber	Schaerer (1987)
1982/11/01	Lefroy Glacier, Lake Louise	Alberta	1 climber	Schaerer (1987)
1982/11/27	Mount Sheer, 36 km north of Vanouver	British Columbia	1 hiker	Schaerer (1987)
1983/01/08	Apex Mountain	British Columbia	1 skier	Schaerer (1987)
1983/02/13	Tent Ridge, Kananaskis Country	Alberta	1 skier	Schaerer (1987)
1983/07/03	Mount Bowlen, Banff National Park	Alberta	1 climber	Schaerer (1987)
1983/09/18	Mount Charlton, Jasper National Park	Alberta	1 climber	Schaerer (1987)
1984/01/07	Quartz Creek near Golden	British Columbia	1 snowmobiler	Schaerer (1987)
1984/01/29	Flathead Pass	British Columbia	1 snowmobiler	Schaerer (1987)
1984/02/11	Redfern Lake	British Columbia	2 snowmobilers	Schaerer (1987)
1984/12/27	Wawa Bowl, Banff National Park	Alberta	1 skier	Jamieson and Geldsetzer (1996)
1984/12/29	Mount Neptune near Rossland	British Columbia	2 skiers	Jamieson and Geldsetzer (1996)
1985/02/18	Monashee Mountains near Blue River	British Columbia	2 skiers	Jamieson and Geldsetzer (1996)
1985/02/23	Onion Mountain near Smithers	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1985/03/02	Mount Erris near Elkford	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1985/03/03	Blanche Mountain	Quebec	1 skier	Jamieson and Geldsetzer (1996)
1986/01/25	White Queen Mountain, Nelson	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1986/02/04	Selkirk Mountains near Nakusp	British Columbia	2 skiers	Jamieson and Geldsetzer (1996)
1986/02/17	Coquihalla Lakes, Cascade Mountains	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1986/03/29	Clemina Creek, Monashee Mountains	British Columbia	4 snowmobilers	Jamieson and Geldsetzer (1996)
1987/03/23	Thunder River, Cariboo Mountains	British Columbia	7 skiers	Jamieson and Geldsetzer (1996)
1987/05/29	Bow Summit, Banff National Park	Alberta	1 skier	Jamieson and Geldsetzer (1996)
1987/06/14	Mount Bryce, Banff National Park	Alberta	3 climbers	Jamieson and Geldsetzer (1996)
1987/08/01	Mount Robson, Mount Robson Provincial Park	British Columbia	1 climber	Jamieson and Geldsetzer (1996)
1988/01/17	Standfast Creek, Selkirk Mountains	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1988/02/07	Crowfoot Pass, Banff National Park	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1988/02/13	Shawinigan	Quebec	1 tobogganer	Jamieson and Geldsetzer (1996)
1988/02/20	Fossil Mountain, Banff National Park	Alberta	2 climbers	Jamieson and Geldsetzer (1996)
1988/03/22	Sale Mountain, Selkirk Mountains	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1988/04/03	Whitetooth Mountain near Golden	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1988/12/28	Extra Light, Yoho National Park	British Columbia	1 ice climber	Jamieson and Geldsetzer (1996)
1989/01/02	Garibaldi Provincial Park	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1989/01/04	Middle Kootenay Pass, Pincher Creek	Alberta	1 snowmobiler	Jamieson and Geldsetzer (1996)
1989/01/28	Telegraph Creek, Coast Mountains	British Columbia	1 resident	Jamieson and Geldsetzer (1996)
1989/03/15	Flute Mountain, Whistler	British Columbia	1 skier	Jamieson and Geldsetzer (1996)

Date (y/m/d)	Location	Province	Fatalities	Source
1989/03/15	Monashee Mountains near Mica Creek	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1990/01/06	Wawa Bowl, Banff National Park	Alberta	1 snowboarder	Jamieson and Geldsetzer (1996)
1990/01/28	Sand Creek, Southern Rockies	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1990/01/30	Kokanee Glacier Provincial Park	British Columbia	2 skiers	Jamieson and Geldsetzer (1996)
1990/02/11	Healy Creek, Banff National Park	Alberta	4 skiers	Jamieson and Geldsetzer (1996)
1990/02/22	Hartly Creek near Golden	British Columbia	1 ice climber	Jamieson and Geldsetzer (1996)
1991/01/20	Rummel Col, Kananaskis Country	Alberta	1 skier	Jamieson and Geldsetzer (1996)
1991/02/05	Malbaie River	Quebec	1 tobogganer	Jamieson and Geldsetzer (1996)
1991/03/12	Bugaboo Creek, Purcell Mountains	British Columbia	9 skiers	Jamieson and Geldsetzer (1996)
1991/04/09	Blackcomb Glacier, Coast Mountains	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1991/11/27	Twin Falls near Smithers	British Columbia	1 ice climber	Jamieson and Geldsetzer (1996)
1992/01/03	Thornhill Mountain near Terrace	British Columbia	2 snowmobilers	Jamieson and Geldsetzer (1996)
1992/02/26	Silk Tassel, Yoho National Park	British Columbia	1 ice climber	Jamieson and Geldsetzer (1996)
1992/04/22	Mount Dagon, Monarch Icefields	British Columbia	2 climbers	Jamieson and Geldsetzer (1996)
1992/12/13	Owl's Head Mountain near Sicamous	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1992/12/19	Mount Strachan near West Vancouver	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1992/12/25	Parker's Ridge, Jasper National Park	Alberta	1 skier	Jamieson and Geldsetzer (1996)
1993/02/18	Tadoussac mountains	Quebec	1 hiker	Jamieson and Geldsetzer (1996)
1993/03/07	Mount Skookum near Carcross	Yukon Territory	1 snowmobiler	Jamieson and Geldsetzer (1996)
1993/03/10	Bourne Glacier near Revelstoke	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1993/03/17	Bruins Pass, Glacier National Park	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1993/03/20	Snow Dome, Jasper National Park	British Columbia	3 climbers	Jamieson and Geldsetzer (1996)
1993/05/24	Hummingbird Ridge, Mount Logan	Yukon Territory	1 climber	Jamieson and Geldsetzer (1996)
1993/08/09	Mount Temple, Banff National Park	Alberta	1 climber	Jamieson and Geldsetzer (1996)
1993/11/28	Mount Howard Douglas, Banff National Park	Alberta	1 skier	Jamieson and Geldsetzer (1996)
1994/01/05	Oscar Creek near Ymir	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1994/02/13	Hasler Creek near Chetwynd	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1994/02/13	Middle Kootenay Pass near Pincher Creek	Alberta	2 snowmobilers	Jamieson and Geldsetzer (1996)
1994/02/22	Greely Creek, Selkirk Mountains	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1994/05/21	Europa Lake, Coast Mountains	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1994/08/31	Mount Athabasca, Jasper National Park	Alberta	1 climber	Jamieson and Geldsetzer (1996)
1994/11/19	Hemlock Valley ski area near Mission	British Columbia	1 ski patroller	Jamieson and Geldsetzer (1996)
1995/02/04	Mount Ryder near McBride	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1995/02/18	100 km southeast of Prince George, Cariboo Mountains	British Columbia	2 tobogganers	Jamieson and Geldsetzer (1996)
1995/02/19	Burstall Pass, Kananaskis Country	Alberta	1 skier	Jamieson and Geldsetzer (1996)
1995/02/24	Cascade Mountain, Banff National Park	Alberta	2 climbers	Jamieson and Geldsetzer (1996)
1995/02/26	Bruce Creek, Purcell Mountains	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1995/03/05	Broadview Mountain, Kakwa recreational area	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1995/03/10	Blanc-Sablon, Quebec	Quebec	2 residents	Jamieson and Geldsetzer (1996)
1995/03/15	Marmot Peak, Jasper National Park	Alberta	1 skier	Jamieson and Geldsetzer (1996)
1995/03/19	Doctor Creek, Purcell Mountains	British Columbia	2 snowmobilers	Jamieson and Geldsetzer (1996)
1995/03/19	Telkwa Range near Houston	British Columbia	1 snowmobiler	Jamieson and Geldsetzer (1996)
1995/11/12	Sawtooth Mountain, Baffin Island	Nunavut	1 snowmobiler	Jamieson and Geldsetzer (1996)
1995/12/22	Robertsonville	Quebec	2 skiers	Jamieson and Geldsetzer (1996)
1996/01/04	No Name Ridge, Stagleap Provincial Park	British Columbia	1 snowboarder	Jamieson and Geldsetzer (1996)
1996/02/26	Kokanee Glacier Provincial Park	British Columbia	1 skier	Jamieson and Geldsetzer (1996)
1996/03/26	Mount Groulx	Quebec	1 skier	Jamieson and Geldsetzer (1996)
1996/05/17	Mount Cerebrus, Coast Mountains	British Columbia	3 climbers	Jamieson and Geldsetzer (1996)
1996/06/05	Mount Logan, Kluane National Park	Yukon Territory	1 climber	Jamieson and Geldsetzer (1996)
1996/12/12	Schaffer Bowl, Yoho National Park	British Columbia	1 skier	Manners, 1997
1996/12/16	Spearhead Mountains near Whistler	British Columbia	3 skiers	Manners, 1997
1997/01/13	Mud River, Monashee Mountains	British Columbia	2 skiers	Manners, 1997
1997/02/02	Lang Creek near Golden	British Columbia	1 snowmobiler	Manners (1997)
1997/03/09	Coquihalla Summit	British Columbia	1 snowboarder	Manners (1997)
1997/03/21	Apex Mountain	British Columbia	1 skier	Manners (1997)
1997/03/22	Mount Jowett near Trout Lake	British Columbia	2 skiers	Manners (1997)
1997/03/30	Lang Creek near Golden	British Columbia	1 snowmobiler	Manners (1997)
1997/03/31	Southwest of Atlin	British Columbia	2 skiers	Manners (1997)

APPENDIX B

Snow avalanche danger scale (provided by the Canadian Avalanche Association)

SNOW AVALANCHE SAFETY BASICS

Snow avalanches don't happen by accident and most human involvement is a matter of choice, not chance. Most avalanche accidents are caused by slab avalanches that are triggered by the victim or a member of the victim's party. However, any avalanche may cause injury or death and even small slides may be dangerous. Hence, always practice safe route-finding skills, be aware of changing conditions, and carry avalanche rescue gear. Learn and apply avalanche terrain analysis and snow stability evaluation techniques to help minimize your risk. Remember that avalanche danger rating levels are only general guidelines. Distinctions between geographic areas, elevations, slope aspects and slope angles are approximate and transition zones between dangers exist.

Canadian avalanche danger descriptors

Danger level (colour) ...WHAT...	**Avalanche probability and avalanche trigger** ...WHY...	**Recommended action in the backcountry** ...WHAT TO DO...
LOW (green)	Natural avalanches very unlikely. Human-triggered avalanches **unlikely**.	Travel is generally safe. Normal caution advised.
MODERATE (yellow)	Natural avalanches unlikely. Human-triggered avalanches **possible**.	Use caution in steeper terrain on certain aspects (defined in accompanying statement).
CONSIDERABLE (orange)	Natural avalanches possible. Human-triggered avalanches **probable**.	Be increasingly cautious in steeper terrain.
HIGH (red)	Natural and human-triggered avalanches **likely**.	Travel in avalanche terrain is not recommended.
EXTREME (red with black border)	Widespread natural or human- triggered avalanches **certain**.	Travel in avalanche terrain should be avoided and travel confined to low-angle terrain well away from avalanche runouts.

Floods

Gregory R. Brooks[1], Stephen G. Evans[1], and John J. Clague[2]

Brooks, G.R., Evans, S.G., and Clague, J.J., 2001: Floods; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 101–143.*

Abstract: In Canada, flooding is caused by hydrometeorological mechanisms individually or in combination, and, less commonly, by the formation and failure of natural dams. Between 1900 and June 1997, there have been 168 known Canadian flood 'disasters'. These events are clustered in the south in Eastern Canada, whereas they generally are spread throughout Western Canada, but with some clustering in southern Manitoba and southwestern British Columbia. Floods are an important part of the flow regime of alluvial rivers, with geomorphic effects ranging from minor/negligible to large-scale erosion and deposition that completely alters the character of the pre-flood channel and valley bottom. Along many rivers, the impacts of flooding have been reduced by the careful management of flood-prone lands and by the construction of flood-protection infrastructure. Climate change may be altering the frequency of extreme flood events and thus increasing the risk of occupying flood-prone lands along some rivers.

Résumé : Au Canada, les inondations sont provoquées par un ou plusieurs mécanismes hydrométéorologiques et, moins souvent, par la formation et la rupture de barrages naturels. Entre 1900 et juin 1997, on a recensé 168 inondations «catastrophiques» au Canada. Alors que dans l'Est du Canada elles sont agglutinées dans le sud, dans l'Ouest canadien, elles se déclenchent un peu partout, mais un certain nombre sont concentrées dans le sud du Manitoba et le sud-ouest de la Colombie-Britannique. Les inondations constituent une importante partie du régime d'écoulement des rivières alluviales et leurs incidences géomorphologiques vont de l'érosion superficielle et négligeable à l'érosion à grande échelle et la formation de dépôts qui modifient complètement les caractéristiques du lit du cours d'eau et le fond des vallées. La gestion prudente des terres inondables le long des rivières et la construction d'infrastructures de défense contre les crues ont permis de réduire les dégâts causés par les inondations. Il se peut que le changement climatique modifie la fréquence des grosses crues et, de ce fait, augmente le risque lié à la construction sur les terrains inondables situés le long de certaines rivières.

---

[1] Terrain Sciences Division, Geological Survey of Canada, Ottawa, Ontario K1A 0E8
[2] Department of Earth Sciences, Simon Fraser University, Burnaby, British Columbia V5A 1S6

INTRODUCTION

Flooding along rivers is part of the natural hydrological cycle and occurs somewhere in Canada every year. In some watersheds, nonfluvial processes also cause flooding both by locally obstructing flow and later by suddenly releasing it. Flooding is a major natural hazard that causes significant disasters in Canada. It has resulted in the loss of over 198 lives in the twentieth century and damage in excess of $2 billion. Worldwide, river (and coastal) flooding affects more people than all other natural hazards.

WHY FLOODING OCCURS

Most commonly, flooding occurs when river discharge temporarily exceeds the capacity of a channel to contain the flow. The water overflows the river banks and inundates adjacent low-lying areas, including the floodplain and, in extreme cases, low terrace surfaces and valleyside margins. At any given location along a river, the extent of flooding depends on the amount of excess water relative to channel capacity — the more flow exceeds channel capacity, the more extensive and/or deeper the inundation of adjacent low-lying areas.

Valley-bottom morphology is another variable important to flooding. For flood flows of similar magnitude, flood levels are higher and narrower along a constricted reach and shallower and broader along a wide valley bottom, assuming the two reaches have similar channel capacity, roughness, and slope (Fig. 1a, b). In settings where a river channel is confined within a deep, narrow gorge or valley, high-magnitude flows are entirely contained and only a part of the valleysides is inundated (Fig. 1c). Because valley-bottom morphology can vary considerably both along a river and between rivers, the extent of flooding will vary accordingly.

Flooding also occurs much more locally where an obstruction blocks flow, causing river waters to pond, rise, and backflood the immediate upstream area (Fig. 2; *see* 'Flood mechanisms' below). The extent of backflooding is a function of the height of the obstruction in the channel or valley bottom and the gradient of the valley bottom upstream of the obstruction. This type of flooding commonly is temporary, as a river usually overtops and erodes the obstruction, or the obstruction fails. Occasionally, however, a permanent lake forms.

FLOOD MECHANISMS

Hydrometeorological floods

Most flooding in Canada is caused by hydrometeorological mechanisms (Table 1) acting either individually or in combination. Hydrometeorological mechanisms are a direct product of climate. The intensity of these mechanisms varies regionally, reflecting the diversity of climate across Canada. Nevertheless, all the mechanisms described below apply broadly to all regions of Canada. Their importance varies considerably throughout the year, again reflecting the connection

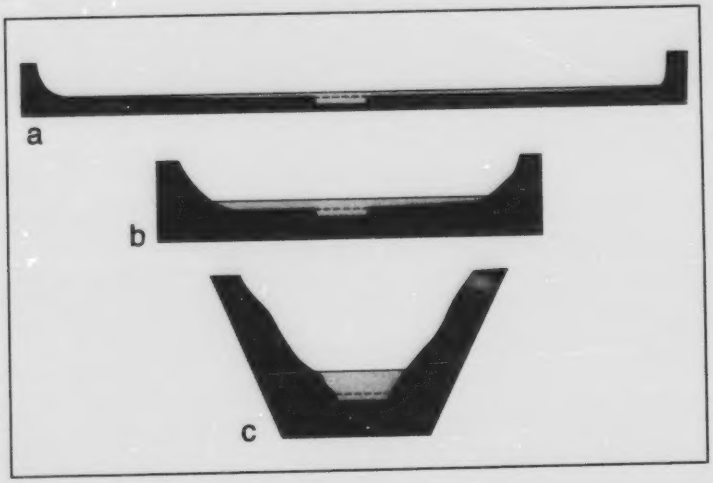
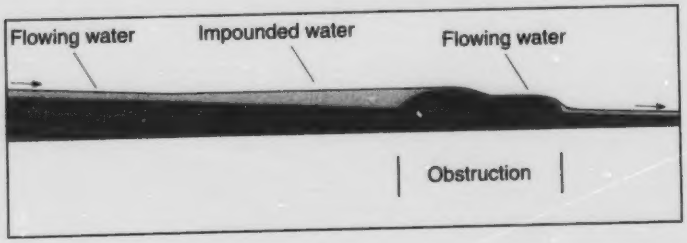


Figure 1.

Variation in the extent of flooding in valleys with different morphologies: ***a****) a valley with a broad floodplain;* ***b****) a valley with a moderately wide floodplain; and* ***c****) a narrow gorge that completely contains the flow. These examples are relatively simple and generalized. In nature, a wide range of valley morphologies exist that have various shapes and sizes relative to the size of the channel, with or without river terraces, and with valley-bottom surfaces composed of alluvial sediments, nonalluvial sediments (e.g. glacial), and/or bedrock.*

Figure 2.

Sketch of localized flooding along a stream reach caused by an obstruction of flow in the channel. Relative to the profile of a moderately sized river, this sketch has a vertical exaggeration of 10 to 20 times.



to climate. Particularly severe flooding may result when several of these mechanisms occur coincidentally, as can happen in the spring.

Snowmelt-runoff floods

During the winter months in most of Canada, precipitation falls as snow and accumulates on the ground. Because of the length of the winter, large quantities of water can be stored as snow. Melting in the spring converts the snow to water, some of which runs off into streams. Flooding occurs when the amount of meltwater runoff exceeds the channel capacity of streams carrying the water away.

The most important factors controlling the amount of snowmelt runoff, and therefore the possibility of flooding, are the thickness, 'ripeness', and extent of the snowpack and the rate of melting. (A 'ripe' snowpack is dense, has a uniform temperature of 0°C, and its liquid-water-holding capacity is saturated; it will release considerably more water at the ground surface than an 'unripe' snowpack, other factors being equal.) The more snow on the ground during a melt period, the more water potentially is available for snowmelt runoff. Ultimately, of course, snowmelt is controlled by available heat energy (radiant and latent energy). Relatively cool weather causes slow melting of the snowpack and the gradual release of meltwater whereas warm weather speeds melting and releases meltwater more rapidly. A winter of high snow accumulation thus does not necessarily produce significant spring flooding.

Table 1. Flood mechanisms.

Hydrometeorological - snowmelt runoff - storm-rainfall - rain-on-snow - ice jam
Natural dams - landslide - glacier - moraine - snow

Additional factors influencing the occurrence of snowmelt flooding are the presence of saturated soil conditions during the previous fall and the development of a deep frost layer prior to the arrival of snow. Both of these conditions limit the infiltration of meltwater into the ground during the subsequent period of snowmelt, causing the released water to flow as surface runoff. However, the presence of warm, dry winds during the melt period will cause the direct evaporation of snow, thus reducing the amount of water that is contributed to runoff.

Severe flooding from snowmelt commonly happens when there is a rapid shift from cold to warm temperatures in the spring following a winter of high snow accumulation. In general, the later the onset of the spring thaw, the more likely that the transition from cold to warm weather will be rapid. Since the climatic factors influencing both the accumulation of snow and the rate of melt are regional, snowmelt flooding commonly occurs over large areas and affects numerous watersheds.

Snowmelt floods are one of the most common types of floods in Canada, occurring in watersheds of all sizes, often in combination with storm-rainfall runoff and/or ice jams (*see* below). For most rivers, the highest average flow (daily, weekly, and monthly) occurs during the snowmelt period, reflecting the importance of snowmelt runoff to river hydrology. This high runoff interval is known as the 'freshet' and usually happens between March and May, but can occur in June or early July in some mountainous or arctic areas. Snowmelt flooding can also occur in the winter, when it is triggered by sudden and rapid thaw. Notable snowmelt floods in the twentieth century occurred in northwestern British Columbia in May–June 1936, along the Fraser River, British Columbia, in May–June 1948, and along the Red River, Manitoba, in May 1950 and May 1997 (Fig. 3; Appendix A).

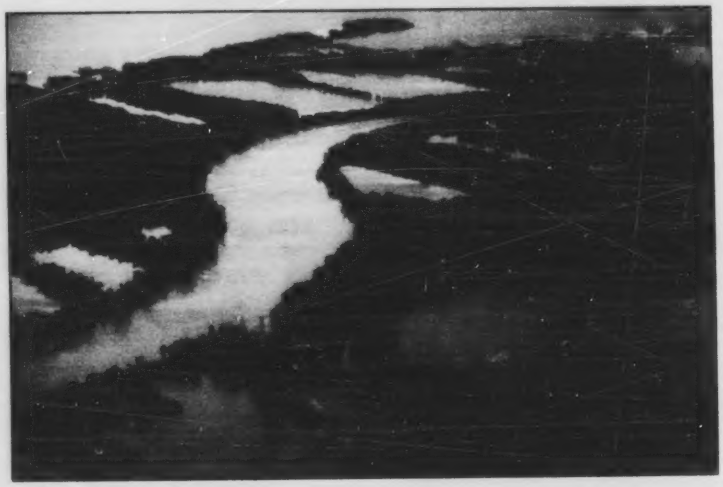


Figure 3.

Flooding along the Red River, Manitoba, 9 May 1997, the result of snowmelt runoff from a heavy winter snowpack. The shallow valley occupied by the Red River has insufficient capacity to contain the flow. Water has overtopped the valleysides and inundated the flat surface of the Red River valley. The river flows from right to left. Photograph by G.R. Brooks. GSC 2000-109

Storm-rainfall floods

Flooding can be caused by heavy or torrential rainfall, even after a period of drought. Such flooding can develop rapidly and unexpectedly, particularly when the ground is saturated, the rate of rainfall far exceeds the capacity of the ground to absorb the water, or a significant portion of the ground surface is covered with impermeable materials, such as concrete or pavement in urban areas. The excess water runs off the land and enters the river system, eventually exceeding the channel capacity of the drainage network.

The magnitude of storm-rainfall flooding depends on the intensity and duration of the rainfall, the areal extent of the storm, pre-storm ground moisture conditions, and drainage basin characteristics (e.g. topography, overburden thickness, vegetative cover, drainage density). Convective storms and mid-latitude and tropical cyclones (including hurricanes) generate these floods. River levels may remain high for periods of hours to several days, depending on the intensity and duration of the storm and the size of the drainage basin.

Figure 4. Flooding in downtown Chicoutimi, Quebec, triggered by heavy rainfall during July 1996. The flood waters from the Chicoutimi River have overtopped a small dam, allowing water to spill through an adjacent urban area. Photograph courtesy of La Presse, ©La Presse

Storm-rainfall floods affect small to large watersheds (up to 100 000 km²) throughout Canada. Rivers draining larger areas generally are unaffected because the zone of intense rainfall within a storm system usually is limited to a relatively small portion of the drainage basin. In such cases, the flood flows of small rivers travel downstream and eventually join the channel of a larger river that is able to contain the flow. An example of this occurred in the Saguenay area, Quebec, in July 1996 when heavy rainfall caused severe flooding along a number of tributaries (drainage basin areas up to several thousand square kilometres) of the Saguenay River (drainage basin area 88 100 km²). Occasionally, however, storm-rainfall flooding will affect very large watersheds, as demonstrated by flooding along the Mississippi River in 1993. In this case, prolonged, heavy rainfall over a large part of the midwestern United States contributed to record flood levels along the main channel of the largest river in North America.

Numerous storm-rainfall floods have occurred in Canada during the twentieth century. Notable examples include the floods in southern Ontario caused by Hurricane Hazel in October 1954, along the Mackenzie River, Northwest Territories, in July 1988, and in the Saguenay area, Quebec, in July 1996 (Fig. 4; Appendix A). More isolated, but particularly severe flooding can be caused by thunderstorms, as for example in Timmins, Ontario, in August 1961, in Edmonton, Alberta, in July, 1978, and in Montréal, Quebec, in July 1987 (Appendix A).

Rain-on-snow floods

Rain-on-snow floods are a combination of snowmelt runoff and storm-rainfall floods. Rainfall runoff is augmented by snowmelt, thus increasing the amount of water available to enter the stream system. Significant factors contributing to the melting of snow during a rainstorm are air and rainfall temperatures, amounts of rainfall, and wind speeds. Snowpack ripeness is also a key variable controlling the amount of surface runoff released as an unripe snowpack can store more rainwater than a ripe snowpack. Depending on the antecedent ripeness of the snowpack, rain-on-snow events having similar meteorological and other snowpack characteristics can generate significantly different amounts of runoff.

When heavy or sustained rainfall occurs in combination with other meteorological conditions to ripen and significantly melt a moderately deep snowpack, rain-on-snow can generate substantial runoff causing very severe flooding. Commonly, the most extreme event recorded along a given river is a rain-on-snow event. Rain-on-snow floods occur in all parts of Canada, but can be particularly severe in the fall along the west coast and during the winter and early spring elsewhere in Canada. Examples include flooding along the St. John River, New Brunswick, in March 1936 and April 1973, in Newfoundland in January 1993, and in southwestern British Columbia during October 1984 (Appendix A).

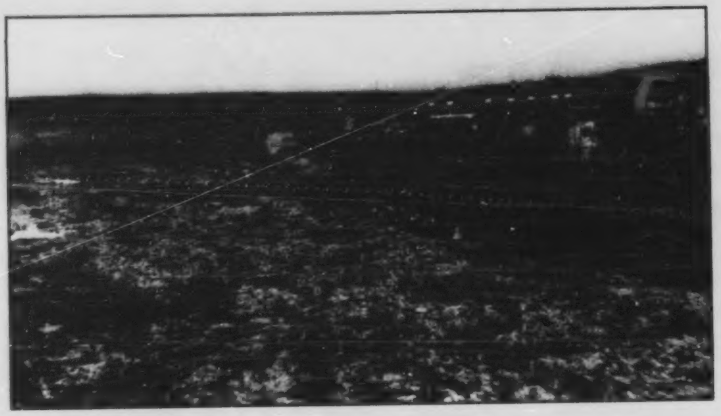


Figure 5.

View upstream of an ice jam along the Mackenzie River at Fort Simpson, Northwest Territories, just downstream from the confluence with the Liard River, 3 May 1989. Here, the ice jam is 1.2 km wide. It is part of a large jam over 30 km long that extends up the Liard River. Note the ponded water between the jammed ice in the foreground and buildings along the river bank. Photograph courtesy of the Water Survey of Canada, Fort Simpson Office

Ice-jam floods

Ice-jam floods result from the temporary obstruction of river flow by the buildup of ice fragments within the channel (Fig. 5). Once formed, an ice jam causes the river to rise immediately upstream. The river may overtop its banks, depending on the height of the obstruction relative to the depth of the channel. Significant ice jams can form during both freeze-up and breakup, but jams formed during breakup have the greater potential for flooding as they commonly coincide with the freshet flow arising from snowmelt. On large rivers, major ice jams can be kilometres long and, in extreme cases, up to several tens of kilometres long (for example, at the confluence of the Mackenzie and Liard rivers; Fig. 5).

This type of flooding is localized because it requires the formation of an ice jam. However, the tendency for an ice jam to form at any given location along a river varies. Specific factors promoting the formation of an ice jam are the presence of local sections of intact ice cover along the river during breakup and/or local channel characteristics including a decrease in channel gradient, channel shoaling, variation in channel width, channel splitting by islands or bars, and sharp bends. An ice jam also can form behind bridges and other artificial structures that constrict a channel. The severity of ice jamming generally increases the greater the thickness of the ice cover, the greater the physical strength of the ice, and/or the greater the supply of flowing (or 'running') ice from upstream. These factors will vary from year to year depending on, for example, the harshness of the winter, the amount of ice decay and melting prior to breakup, and the amount of rise in river level immediate prior to and at the time of breakup.

Reaches downstream from an ice jam also can be affected by flooding. Ice jams can suddenly fail or move rapidly, generating a surge of water and ice that increases river level downstream quickly. The magnitude of the surge depends partly on the size of the ice jam and the volume of obstructed water. A surge from a large ice jam can flood low-lying areas downstream.

Ice jamming is a significant flood mechanism along rivers in most of Canada. It is not generally a factor in areas where a moderate winter climate prevents the buildup of river ice (e.g. the southwest coast of British Columbia). Although the peak discharge along rivers where ice jamming is significant results from other processes (snowmelt, storm-rainfall, or rain-on-snow runoff), the high water level — the direct cause of flooding — often is a product of ice jamming. For example, a study of nineteenth- and twentieth-century flooding at communities along the Mackenzie River, Northwest Territories, linked the vast majority of flood events to ice jams during spring breakup. Large rivers that undergo significant flooding from ice jamming are the Mackenzie (Fig. 5), Red (Manitoba), Saint John (New Brunswick), and St. Lawrence (Ontario–Quebec) rivers.

Flooding related to natural dams

The formation and failure of natural dams also cause flooding, although it is more localized and less common than hydrometeorological flooding. Floods from natural dams involve the blockage of drainage by landslides, glaciers, moraines, and snow (Table 1). Flooding occurs upstream of the dam as a result of ponding and downstream due to the sudden release, or 'outburst', of the impounded water when the dam fails. Such 'outburst floods' can have peak discharges one order of magnitude or more greater than the maximum expected 'normal' hydrological flood and may cause enormous damage along their flood paths.

Landslides dams

Landslide dams form where landslide debris blocks drainage on a valley floor. The scale of damming depends upon the thickness and extent of the debris on the valley bottom. Large landslides consisting of several hundred million cubic metres of debris can form significant lakes within major valleys (Fig. 6). Immediately after a landslide dam forms, flooding begins upstream as river waters pond behind the landslide debris. The rate of water rise is a function of the discharge of the river at the time of the landslide relative to the volume of the impoundment basin. Natural reservoirs can take hours to many weeks to fill.

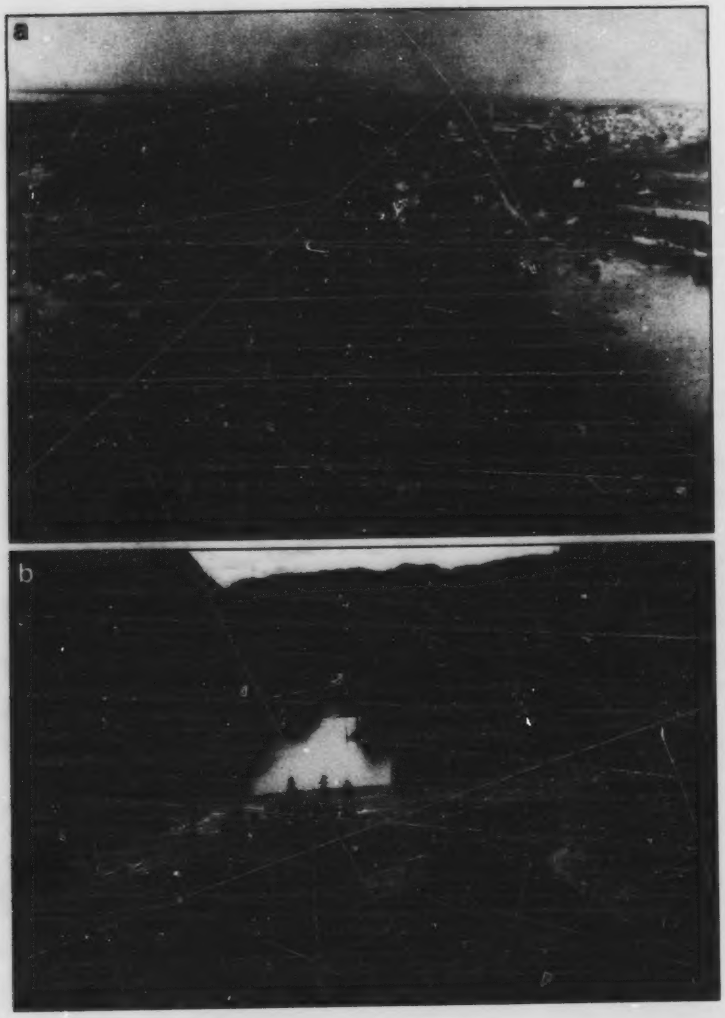


Figure 6.

Landslide-dammed rivers. ***a)*** *Oblique aerial view of flooding at Rimouski, Quebec, caused by the temporary blockage of the Rimouski River by an earthflow in leda clay (L) in August 1951; view northwest. Photograph courtesy of J.E. Hurtubise.* ***b)*** *Landslide dam on Meager Creek, Mount Meager Volcanic Complex, southern Coast Mountains, British Columbia. Debris deposited by several debris-flow surges from Capricorn Creek (at right) formed a fan (foreground) that blocked Meager Creek in July 1998. The dam created a 2 km long lake that persisted for at least two months. Photograph by S.G. Evans, taken 15 July 1998. GSC 2000-110*

Most landslide dams are temporary and last only days to months after overtopping, depending on the size and shape of the dam and the composition of the debris. Overtopping waters will incise the obstructing debris, causing the impoundment to drain. Erosion commonly begins at the downstream toe of the debris and migrates upstream to the crest, whereupon rapid incision may ensue, breaching the impoundment and generating an outburst flood. Landslide dams consisting of blocky rock debris (e.g. from rock avalanches originating in granitic rocks) are comparatively resistant to rapid erosion, and some lakes impounded by these materials become permanent features of the landscape (Fig. 7). Landslide dams consisting of Quaternary silt and clay (e.g. glaciolacustrine or glaciomarine sediments) commonly erode relatively slowly after overtopping, gradually releasing the impounded water.

Flooding related to landslide dams occurs in several parts of Canada (Fig. 8; Table 2). It is most common in the Cordillera where numerous landslides and narrow, steep-sided valleys create conditions favourable for landslide damming. In the Plains region, landslide dams have temporarily blocked major rivers, for example, the Peace River. In Eastern Canada, landslides in 'leda clay' (glaciomarine silt and clay sediments) commonly block rivers and flood low-lying areas upstream (Fig. 6). Catastrophic outburst floods have resulted from the failure of landslide dams formed by large debris avalanches from Quaternary volcanoes in southwestern British Columbia, from the failure of landslide dams in steep mountain valleys elsewhere in the Cordillera, and in one case in Eastern Canada, from the overtopping of a major landslide in leda clay near Saint-Alban, Quebec, in 1894.

Settlements and infrastructure in both Eastern and Western Canada have been temporarily inundated from backflooding caused by landslide damming. In 1840, for example, a large landslide in leda clay on the Maskinongé River, Quebec, blocked the river and formed a lake that extended about 14 km upstream and flooded numerous houses. The dam breached two days later. In 1880, a landslide blocked the Thompson River, south of Ashcroft, British Columbia, impounding a lake about 8 km long that flooded fields and farm buildings upstream. The dam blocked the flow of the Thompson River for 44 hours before it was overtopped and incised. Landslide dams can have a severe impact on fisheries by forming rapids impassable to spawning fish, or unleashing an outburst flood that deeply erodes spawning beds. In British Columbia, landslide dams may have affected prehistoric aboriginal cultures by blocking Pacific salmon runs, for example in the Lillooet area along the Fraser River.

Figure 7. Oblique aerial view of a permanent landslide-dammed lake on the Kennedy River, Vancouver Island, British Columbia. The lake was formed in 1970 by rock-avalanche debris consisting of blocky granodiorite. Note the dead trees killed by rising lake waters behind the dam (lower left). Photograph by S.G. Evans, taken in 1991. GSC 2000-111

Figure 8. Map showing locations of natural dams that are known to have formed and/or failed in the periods 1750–1997 (Eastern Canada) and 1850–1997 (Western Canada) (data from *Tables 2, 3, and 4)*

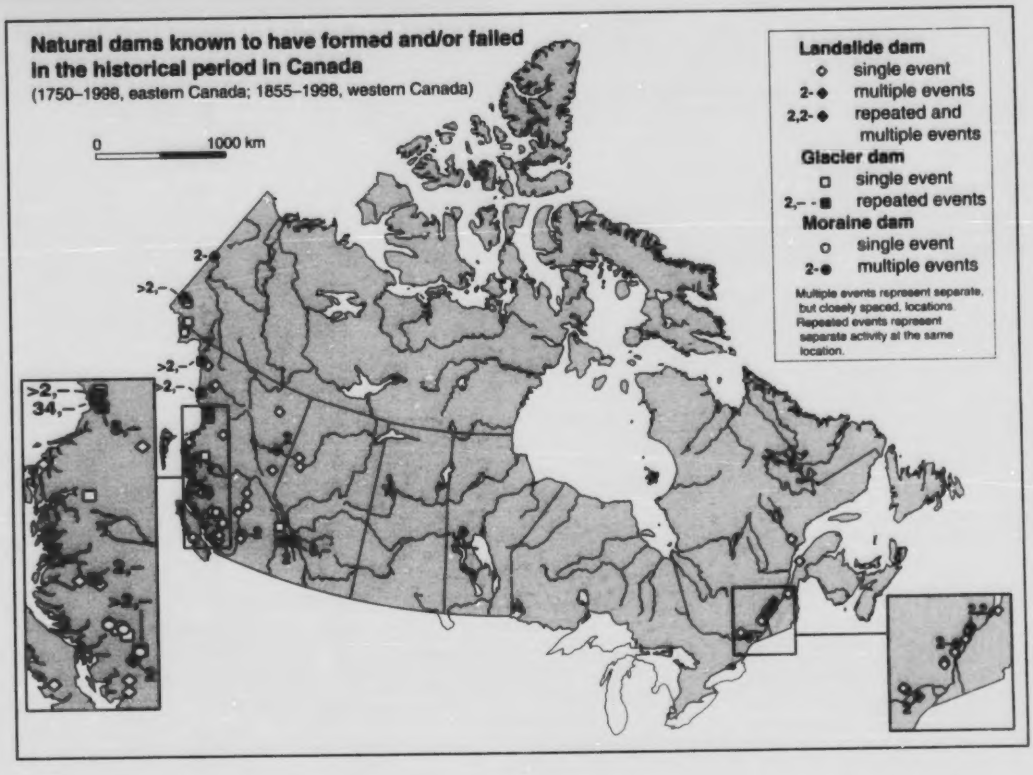

Natural dams known to have formed and/or failed in the historical period in Canada
(1750–1998, eastern Canada; 1855–1998, western Canada)
0
1000 km
Landslide dam
◇ single event
2- ◆ multiple events
2,2- ◆ repeated and multiple events
Glacier dam
□ single event
2,– - ■ repeated events
Moraine dam
○ single event
2- ● multiple events
Multiple events represent separate, but closely spaced, locations. Repeated events represent separate activity at the same location.
2-
>2,–
>2,–
>2,–
>2,–
34,–
2,–
>2,–
2
2,2
2
2

Table 2. Landslide dams known to have formed or failed in the periods 1750–1997 (Eastern Canada) and 1850–1997 (Western Canada).

Site name	Province	Dammed river	Date
Eastern Canada			
Saint-Pierre	Quebec	Sud River	1771
Hayotte	Quebec	Champlain River	1823
Maskinongé River	Quebec	Maskinongé River	1840
Caxton	Quebec	Machiche River	1883
Saint-Alban North	Quebec	Sainte-Anne-de-la-Pérade River	1894
Saint-Alban South	Quebec	Sainte-Anne-de-la-Pérade River	1894
Saint-Alban	Quebec	Sainte-Anne-de-la-Pérade River	1897
Saint-Thuribe	Quebec	Blanche River	1898
Poupore	Quebec	Lièvre River	1903
Notre-Dame-de-la-Salette	Quebec	Lièvre River	1908
Rimouski	Quebec	Rimouski River	1951
Toulnustouc River	Quebec	Toulnustouc River	1962
South Nation River	Ontario	South Nation River	1971
Saint-Ambroise-de-Kildare	Quebec	Blanche River	1975
Saint-Liguori	Quebec	Rouge River	1989
Lemieux	Ontario	South Nation River	1993
Saint-Boniface-de-Shawinigan	Quebec	Machiche River	1996
Western Canada			
Black Canyon	British Columbia	Thompson River	1880
Spences Bridge	British Columbia	Thompson River	1880
North Pacific Cannery	British Columbia	unnamed creek	1891
Spences Bridge	British Columbia	Thompson River	1899
Spences Bridge	British Columbia	Thompson River	1905
Basque Ranch	British Columbia	Thompson River	1921
Big Bend (Quesnel)	British Columbia	Fraser River	1921
Devastator Creek	British Columbia	Meager Creek	1931
Montagneuse River	Alberta	Montagneuse River	1939
South Forks Canyon	British Columbia	Mess Creek	1947
Babine	British Columbia	Babine River	1951
Cheekye River	British Columbia	Cheakamus River	1958
Asseek River	British Columbia	Asseek River	1960s
Farwell Canyon	British Columbia	Chilcotin River	1964
Tahltan River	British Columbia	Tahltan River	1964
Clinton Creek	Yukon Territory	Clinton Creek	1970s
Kennedy River	British Columbia	Kennedy River	1970
Klattasine Creek	British Columbia	Homathko River	between 1971 and 1973
Attachie	British Columbia	Peace River	1973
Wolverine Creek	Yukon Territory	Wolverine Creek	1974
Devastator Creek	British Columbia	Meager Creek	1975
Inklin River	British Columbia	Inklin River	1979
Talchako	British Columbia	Talchako River	1980s
Turbid Creek	British Columbia	Squamish River	1984
Halfway River	British Columbia	Halfway River	1989
Murray River	British Columbia	Murray River	1990
Prophet River	British Columbia	Prophet River	1990
Rycroft	Alberta	Saddle River	1990

Glacier dams

Lakes form when glaciers block drainage as they either advance or retreat. Most glacier-dammed lakes form in either a tributary valley dammed by a main valley glacier or a main valley blocked by a tributary glacier (Fig. 9). They also develop supraglacially, subglacially, between coalesced glaciers, and between a glacier and a valleyside (Fig. 9). Most glacier-dammed lakes in Canada are found in high mountain valleys in the Cordillera (Fig. 8; Table 3). Some of these lakes form during glacier surges, but most border retreating glaciers. Large glacier-dammed lakes are present in Canada's Arctic Archipelago, but have not been studied in detail.

Glacier dams may fail catastrophically, creating an outburst flood or 'jökulhlaup' (an Icelandic word describing a sudden glacier flood; Fig. 10). Two types of failure mechanisms exist, 1) the formation and enlargement of a tunnel (or tunnels) within or at the base of a glacier and 2) the development of a breach from the mechanical failure of part of the ice dam (e.g. the wholesale failure of the dam or collapse of the tunnel roof). The latter process has not yet been documented in Canada, but has occurred elsewhere where glaciers surged across a main valley or an existing body of water (a fiord or lake).

The initiation of lake drainage is a function of a delicate interaction between the length of the drainage path through the glacier, the thickness of the glacier dam, and the hydrostatic head between the lake surface and the drainage outlet. Once drainage begins, heat from the flowing water melts the enclosing ice, progressively enlarging the tunnel system. This in turn increases the flow of escaping water, thus generating more heat and further enlarging the tunnel. This 'positive feedback' process of tunnel enlargement causes an exponential increase in discharge over a period of hours or days, creating a flood. Outflow continues until the lake completely drains or, more commonly, until the level of the lake drops below the level of the outlet tunnel. Glacier-dam failures occur more commonly towards the end of the melt season in late summer or fall when lake levels are high and the subglacial drainage network is best developed.

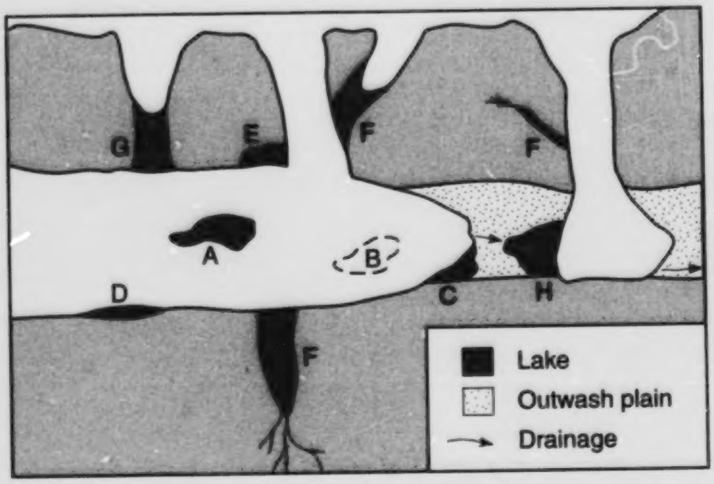


Figure 9.

Schematic diagram showing locations where glacier-dammed lakes can form. ***A****, supraglacial;* ***B****, subglacial;* ***C****, proglacial;* ***D****, embayment at valley-side along glacial margin;* ***E****, area of coalescence of two glaciers;* ***F****, tributary valley adjacent to a trunk or tributary glacier;* ***G****, same as F except glaciers dam both ends of lake;* ***H****, main valley adjacent to a tributary glacier* (after *Clague and Evans, 1994, Fig. 24*).

Table 3. Glacier dams known to have failed in the period 1800–1997 (Western Canada) (*after* Mathews and Clague, 1993; Clague and Evans, 1994).

Site (Lake)	Damming glacier	Province	Date	Comment
Tide	Frank Mackie	British Columbia	1800	- several large jökulhlaups in the nineteenth and early twentieth centuries
Donjek	Donjek	Yukon Territory	ca. 1810	- single event?
Alsek	Lowell	Yukon Territory	ca. 1850	- single event?
Tulsequah	Tulsequah	British Columbia	between 1910 and 1990	- multiple jökulhlaups in this period (commonly one a year in recent decades)
Bridge	Bridge	British Columbia	between 1935 and 1947	- probably several jökulhlaups in this period
Strohn	Bear	British Columbia	1958, 1959, 1960, 1961, and 1962	- five known events
Summit	Salmon	British Columbia	1961, 1965 to 1997	- 34 jökulhlaups, including three floods thought to be jökulhlaups; one event a year occurred between 1965 and 1997
Hazard	Steele	Yukon Territory	1978	- age of most recent jökulhlaup; multiple events occurred at this site prior to 1978
Flood	Flood	British Columbia	1979	- age of most recent jökulhlaup; multiple events occurred at this site prior to 1979
Bigmouth	unnamed glacier	British Columbia	1980s	- single event
MT	Queen Bess	British Columbia	1982	- single event
Ape	Fyles	British Columbia	1984, 1986	- two events
Cathedral	Cathedral	British Columbia	1925, 1946, 1962, 1978, 1982, and 1984	- six known events
Brucejack	Knipple	British Columbia	1989	- single event
Kitimat	unnamed glacier	British Columbia	1993	- single event
Goddard	Goddard	British Columbia	1994	- single event

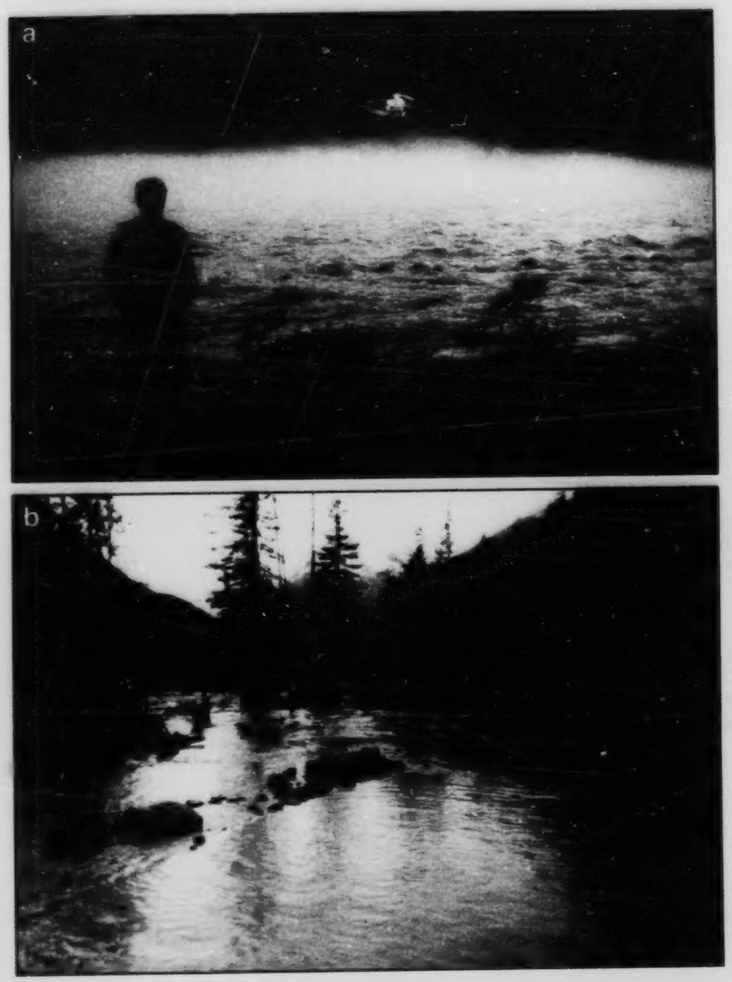


Figure 10.

Jökulhlaup floods in progress. ***a)*** *Jökulhlaup from Summit Lake, near Ninemile, Alaska, 29 July 1993. Floodwaters cover the full breadth of the Salmon River valley, about 1 km wide. The peak discharge was about 1500 m^3/s, which is 50 times larger than the mean discharge of the Salmon River. Photograph by J.J. Clague. GSC 2000-112* ***b)*** *Jökulhlaup from Ape Lake, Noeick River valley, British Columbia, 20 October 1984, about 12 km downstream of Fyles Glacier, which formed the ice dam. The photograph shows the margin of the flood flow where it inundates a logging road. The main flow is visible behind the trees to the left of centre; note the standing waves. Discharge is estimated to have peaked at about 1155 m^3/s at this location. Photograph courtesy of J.R. Desloges.*

Once the outburst flood ceases, the lake can re-form if the tunnel closes or, in the case of a breach, if the glacier readvances to again block the drainage. A re-formed glacial lake can drain again at some time in the future. For example, drainage at Ape Lake, British Columbia, occurred on two separate occasions (Table 3; Fig. 11a), whereas drainage at Summit Lake, British Columbia (Table 3; Fig. 11b), has occurred nearly annually since 1961 (to at least 1997 at the time of writing). The sequence of repeated lake filling and draining will eventually cease when either the glacier retreats to where it no longer impounds drainage and the lake cannot re-form, or the glacier readvances and forms a stable dam. At present, the latter scenario is much less likely to occur as glaciers in Canada have generally been retreating since the late nineteenth century in response to climate warming.

Although most jökulhlaups in Canada occur in remote areas (Table 3), they pose a serious threat to infrastructure, forest lands, and fish habitat far downstream of the glacier dam. Because glaciers are located in remote areas, inundation upstream of a glacier dam has not affected any permanent settlement in Canada. However, the present site of Haines Junction, Yukon Territory, was inundated in the mid-nineteenth century by a glacier-dammed lake impounded behind Lowell Glacier.

Moraine dams

Glaciers in the Cordillera reached their maximum postglacial extent in the period known as the 'Little Ice Age', which spanned several hundred years ending in the nineteenth century. Since the late nineteenth century, glaciers have retreated and thinned in response to climate warming. Moraine-dammed lakes have formed in basins vacated by retreating glaciers, typically in cirques and steeped-walled valleys. A moraine dam may consist of a single moraine or a series of nested moraines. They generally are steep sided and have a relatively low width-to-height ratio and heights in the order of a few tens of metres. Moraines are generally composed of poorly sorted, stratified to massive, blocky or bouldery deposits with a matrix of sand and gravel. Some moraines are cored by relict glacier ice and/or have interstitial ice in the pore space of the sediments.

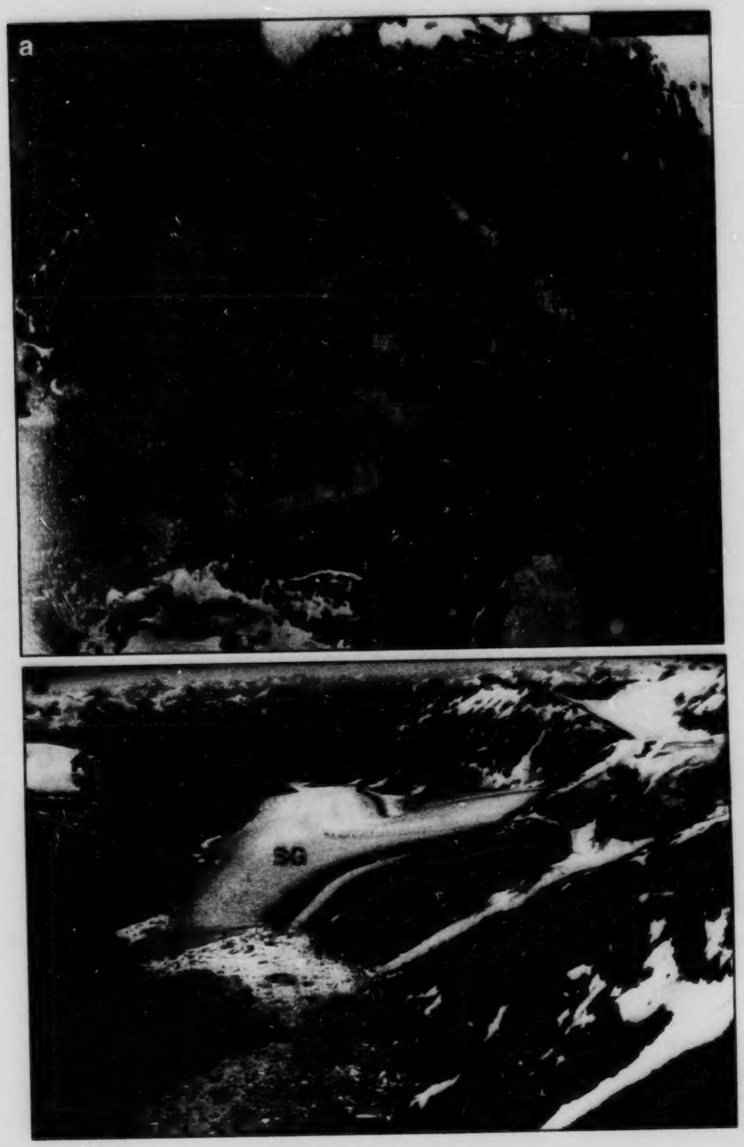


Figure 11.

***a)** Vertical aerial photograph of the maximum extent of Ape Lake dammed by Fyles Glacier in the Coast Mountains, British Columbia; photograph taken in 1978. Ape Lake was partly emptied by jökulhlaups in 1984 and 1986, and has not refilled to date. British Columbia Government air photo BC78125: 019 used with permission from Geographic Data BC, Ministry of Environment, Lands and Parks. **b)** Oblique aerial view of Summit Lake (SL), dammed by the Salmon Glacier (SG) in the Coast Mountains, British Columbia, viewed downvalley. Summit Lake first drained in 1961 and has drained and refilled nearly every year since 1961. During a jökulhlaup, the escaping water travels 11 km at the base of Salmon Glacier before emerging in the Salmon River valley downstream. Photograph by S.G. Evans, taken in 1989. GSC 2000-113*

Some moraine dams are vulnerable to rapid erosion by waters flowing from the lake, which may lead to a catastrophic outburst flood. Failure may result from a number of causes, including an icefall or landslide into the lake, which produces waves that overtop and initiate incision of the outlet (Fig. 12), piping, or settlement of the moraine from the melting of buried ice. Once initiated, erosion of the dam commonly continues until the moraine is completely breached and the lake is emptied. In some cases, erosion is arrested by the development of a coarse lag that armours the outlet channel or by the exposure of a bedrock sill at the outlet as a result of channel incision. The failure of a moraine-dammed lake is a one-time event at least in the short term, as the moraine cannot be rebuilt without a readvance of the glacier that originally built it.

Since 1925, ten outburst floods from moraine-dammed lakes have been documented in the Cordillera (Fig. 8; Table 4). The largest occurred in the Nostetuko River basin in the southern Coast Mountains, British Columbia. In 1983, approximately 6 x 10^6 m^3 of water were released when the moraine dam impounding Nostetuko Lake failed and in 1997, a similar volume was released from nearby Iceberg Lake. In both instances, glacier ice avalanched into the lakes, generating wave trains that overtopped and triggered rapid erosion of the moraine dams (Fig. 12). The ensuing outburst floods devastated the valley bottom many tens of kilometres downstream from the moraines.

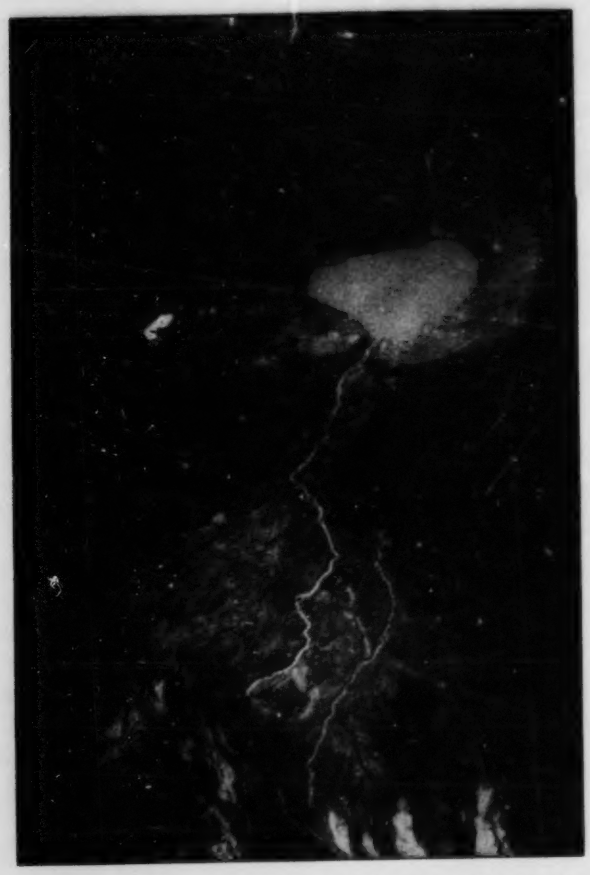


Figure 12. Oblique aerial view of the breached Little Ice Age moraine at Nostetuko Lake, Coast Mountains, British Columbia. In 1983, ice avalanching from the retreating toe of Cumberland Glacier (CG) generated waves in Nostetuko Lake that led to the breaching of the moraine and the sudden release of 6 x 10^6 m^3 of water. Over 10^6 m^3 of material were eroded from the moraine dam. Photograph by S.G. Evans, taken in 1987. GSC 2000-119

Snow

Thick accumulations of avalanched or windblown snow can block streams, forming small, temporary lakes that may subsequently drain suddenly to produce a flood. Flooding resulting from snow dams is not well documented in Canada and is thought to be of only local significance. An example of a snow-dam flood occurred in July 1979 at Tingmisut Lake, Melville Island, Nunavut, and resulted in the drainage of 1.6 x 10^6 m^3 of water in 36 hours from a very small watershed.

FLOOD DISASTERS IN CANADA

An indication of the flood hazard in Canada can be obtained from the occurrence of 'damaging' floods. Although a comprehensive national listing of damaging floods in the twentieth century does not exist, there are lists for some provinces (Nova Scotia, New Brunswick, Newfoundland, northwestern British Columbia, Ontario). A list of the larger damaging Canadian floods in the twentieth century is contained in the Emergency Preparedness Canada disaster database. Emergency Preparedness Canada defines a disaster as follows:

> ...an interruption in time and space of normal processes causing death, injury or homelessness, economic or property loss, and/or significant environmental damage. The interruption is beyond the coping capacity of the community and/or is beyond the assumed risks of human activity. Assumed risk is inherent in most human activity such as transportation and handling of dangerous goods. The interruption precludes war. (Tudor, 1997)

Flood 'disasters' extracted from this database are listed in Appendix A and mapped in Figure 13.

Appendix A includes 168 flood disasters occurring between 1900 and June 1997. This is not by any means a list of flood 'events' as the vast majority of floods in Canada have not been disasters. The data in Appendix A are also biased towards the more densely populated areas of Canada where floods are more likely to have an impact on humans (Fig. 13). Despite these limitations, the database provides an indication of the significance, impact, and location of damaging floods in Canada.

Table 4. Moraine dams known to have failed in the period 1850–1997 (Western Canada).

Site (Lake)	Province	Watershed	Date
Tide	British Columbia	Bowser River	between 1927 and1930
South Mount Macoun	British Columbia	Beaver River	before 1949
Patience Mountain	British Columbia	Jumping Creek	between 1951 and 1956
Big Lake	British Columbia	Big Lake Creek	1960s
Bridge	British Columbia	Bridge River	between 1964 and 1970
Klattasine	British Columbia	Homathko River	between 1971 and 1973
North Mount Macoun	British Columbia	Beaver River	1983
Nostetuko	British Columbia	Homathko River	1983
Tats	British Columbia	Tats Creek	1990
Iceberg	British Columbia	Homathko River	1997

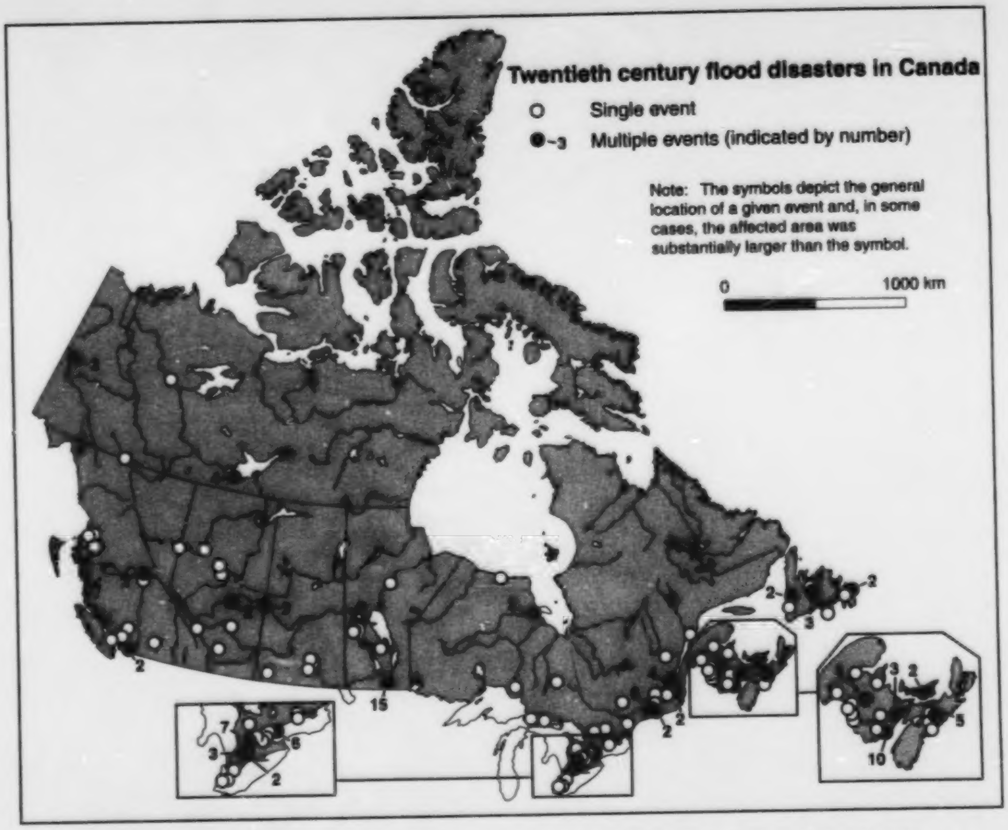

Twentieth century flood disasters in Canada
○ Single event
●-3 Multiple events (indicated by number)
Note: The symbols depict the general location of a given event and, in some cases, the affected area was substantially larger than the symbol.
0
1000 km
2
15
7
3
6
2
2
2
2
2
3
3
2
5
10

Figure 13. Map showing the general location of twentieth-century flood disasters in Canada (data from *Appendix A).*

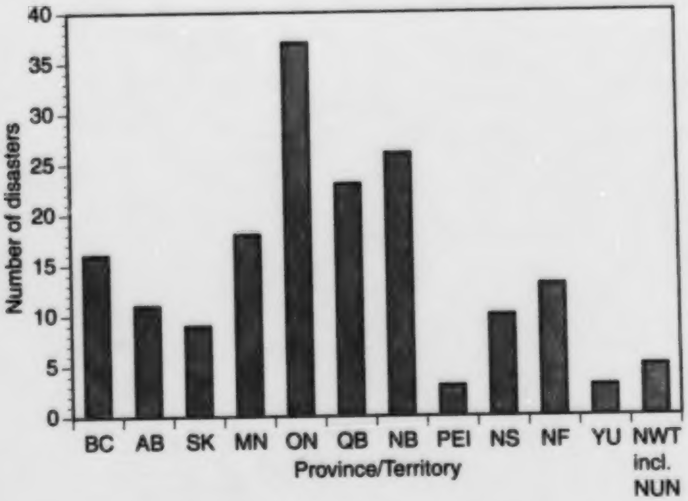


Figure 14.

Distribution of twentieth-century flood disasters in Canada by province and territory (data from *Appendix A).*

Flood disasters have occurred in all Canadian provinces, in the Yukon Territory and in the Northwest Territories (Fig. 13, 14). In Eastern Canada, flood disasters are clustered in the south, reflecting the concentration of population in this area. The distribution is much more diffuse in Western Canada, although some clustering is found in southern Manitoba and southwestern and northwestern British Columbia. About 62% of the disasters listed in Appendix A occurred in Ontario (37 events), New Brunswick (26 events), Quebec (23 events), and Manitoba (18 events; Fig. 103). The large number of disasters in New Brunswick is striking because of the small size of the province relative to the other provinces and undoubtedly reflects a relatively high level of development on flood-prone lands. Specific areas that stand out because of their recurrent flood problems are the Saint John River basin, New Brunswick, (16 disasters), and the Red River basin, Manitoba (15 disasters, including the Assiniboine River). The relatively few disasters in the Northwest Territories (five events), including what is now Nunavut, and the Yukon Territory (three events), which represent about 40% of the area of Canada, reflect the very sparse population in the North. Certainly, some northern communities have repeated flooding problems (e.g. Aklavik), but the local residents generally are knowledgeable about flooding hazards and are able to cope.

During the twentieth century in Canada, flooding has directly and indirectly caused at least 198 deaths and over $2 billion in damage. This figure of 198 deaths must be considered a minimum because damaging floods not included in the database have almost certainly claimed additional lives. In terms of loss of life, the greatest Canadian flood disaster by far was caused by Hurricane Hazel, which struck southern Ontario in October 1954, killing 81 people (Appendix A).

The damage costs shown in Appendix A are difficult to compare because they relate to the year of occurrence and have not been corrected for inflation. The most costly flood in Canada was the May 1950 flood disaster in the Red River Valley, Manitoba, the costs of which are estimated to be $1.09 billion (1999 dollars).

The accounts of many flood disasters in the Emergency Preparedness Canada database are too vague to identify the flood mechanisms positively. However, a study of the descriptions reveals that over 65% of the flood disasters are the result of snowmelt runoff, storm rainfall, or rain-on-snow (combinations of snowmelt runoff and storm rainfall are inferred to be rain-on-snow). Hurricanes, ice jams, storm-rainfall–snowmelt-runoff–ice-jam combinations, and snowmelt-runoff–ice-jam combinations are flood mechanisms of secondary importance. Hurricanes are significant in the Maritime Provinces because they track northwest along and off the eastern seaboard of North America. Only one flood in the disaster database is the result of a natural dam failure (Kicking Horse Pass, British Columbia, 7 September 1978; Appendix A).

Flood disasters occur throughout the year, although most (about 40%) happen in April and May, which is the period of snowmelt throughout much of Canada (Fig. 15). Spring is also the time when several common flood mechanisms (snowmelt runoff, storm rainfall, and ice jams) are likely to occur together, thus increasing the likelihood of high flows. Many of the flood disasters in the January to March period are the result of rain-on-snow during winter mild spells. Floods in the June to November period are the result of severe or prolonged rainstorms. Interestingly, 11 of the 20 flood disasters in Eastern Canada (Ontario to Newfoundland) in August, September, and October were caused by hurricanes or their remnants. Flood disasters are least numerous in November and December.

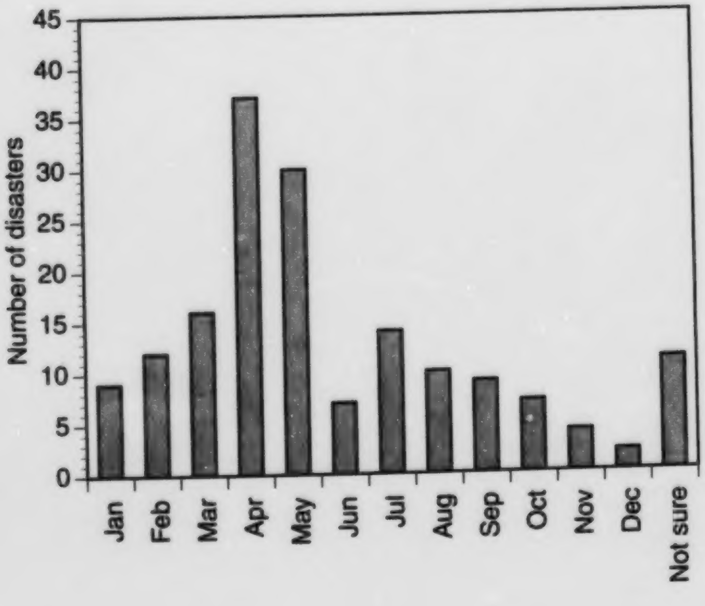


Figure 15.

Distribution of twentieth-century flood disasters in Canada by month (data from *Appendix A*).

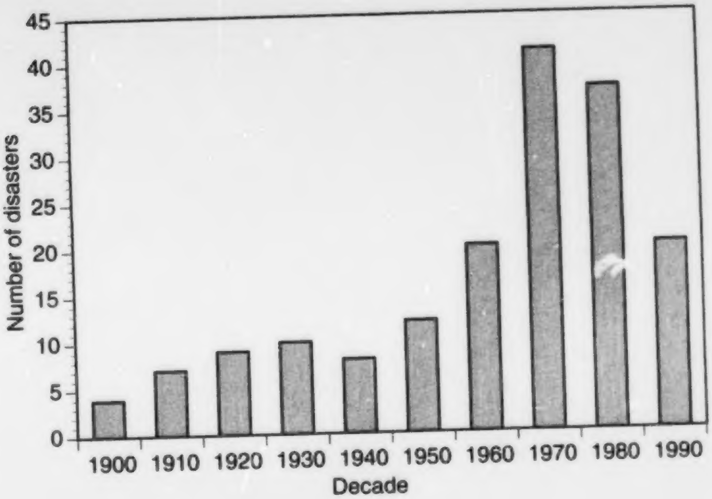


Figure 16.

*Distribution of twentieth-century flood disasters in Canada by decade (*data from *Appendix A).*

The data in Appendix A suggest that the number of flood disasters has increased through the twentieth century, with about 70% occurring after 1959 (Fig. 16). This trend likely reflects several factors. Recent research indicates that the magnitude of flood events in the second half of the twentieth century has generally increased relative to the first half, at least along some rivers. This is thought to reflect a shift in climate. Also, over the twentieth century, development has increased on flood-prone lands as Canada's population has grown. Finally, reporting of flood disasters has improved over the past several decades. Relating to the last point, smaller pre-1960 flood disasters probably are under-represented in the database because they were reported mainly in small, local newspapers and the details provided may have been scant.

EFFECTS OF FLOODING

Geomorphic effects

Along alluvial channels, flood flows represent the upper portion of the range of 'high' flows (including near-bankfull, bankfull, and over-bankfull (i.e. flood) discharges) that cause significant sediment transport, because the velocity of flow at any given location within the channel increases as the stage of flow rises. Eventually a velocity is reached somewhere within the channel whereby the flow can begin to entrain sediment that will be transported and deposited downstream. High flows thus cause cutbank erosion, point-bar accretion, mid-channel bar shifting, lateral channel migration, and/or avulsions, depending on the inherent erosional and depositional processes of a given reach of alluvial channel. The intensity of these processes can increase at least locally during floods, particularly large floods, but along many channel reaches the scale of erosion and deposition from flood flows is not significantly different from that produced by smaller near-bankfull or bankfull flows. In some cases, the geomorphic effects of flooding can be minor and channel morphology is essentially unchanged by the flood flow (Fig. 17).

In some instances, however, large floods cause large-scale erosion and deposition, completely altering the character of the pre-flood river and valley bottom. Erosion can result in significant channel widening, channel avulsions, a decrease in sinuosity, and in some cases, incision. In such cases, once a certain discharge level is surpassed, the overall energy (or 'stream power') of the flow exceeds the erosive threshold of the valley bottom, leading to large-scale erosion of both the channel perimeter and the floodplain (Fig. 18a). The erosive threshold relates to factors such as the hydraulic roughness, vegetative cover, morphology, and material composition of the valley bottom. Thus, the erosive threshold varies between both rivers and river reaches.

In addition to discharge, the stream power of a flood flow is directly proportional to river gradient and inversely proportional to flow width. In cases where the river gradient is very low, the stream power of a flow will not exceed the erosive threshold under any plausible floods and flooding will not cause large-scale erosion of the valley bottom. However, if stream power is sufficiently high, large floods also can lead to erosion along bedrock reaches or reaches composed of very coarse-grained or cohesive, fine-grained sediments that normally undergo negligible or very minor erosion during lesser flows. This may cause, for example, vegetation and surficial sediments (alluvium and other sediments) to be stripped away along the flood path (Fig. 18b).

Once the erosive threshold is exceeded, the amount of erosion that occurs depends on the duration of the flow above the threshold and on the recent flood history of the river. Significant channel widening during a flood represents an adjustment in the morphology of the river to accommodate an extreme discharge and dissipate flow energy. Over a period of years following the flood, the channel will gradually revert to its general pre-flood morphology through sedimentation and/or

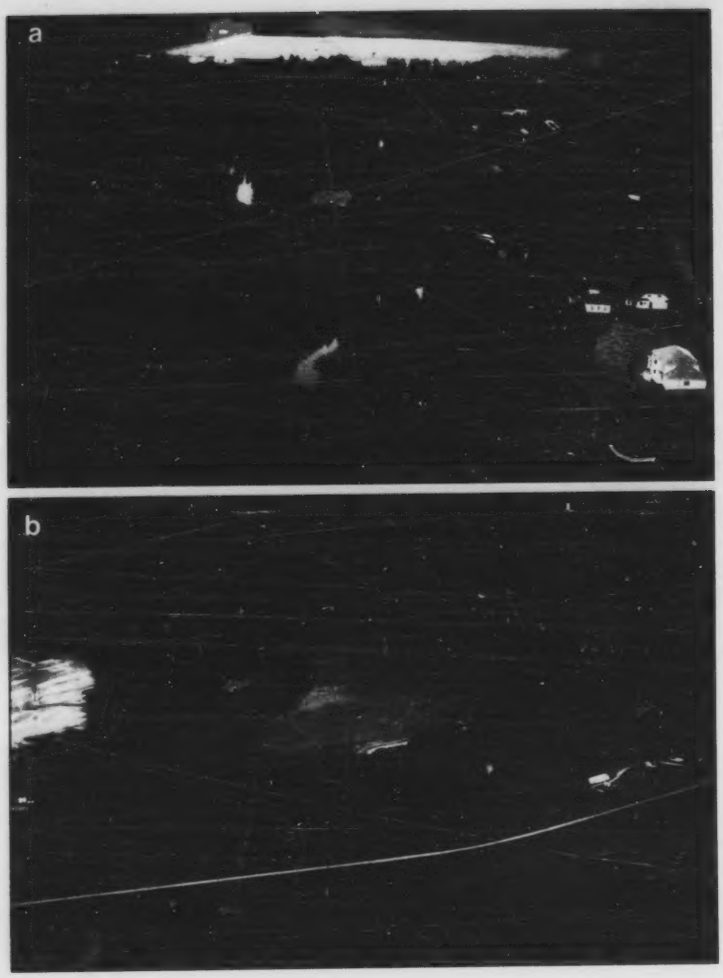


Figure 17.

Examples of river channels in which negligible to minor erosion and deposition occurred during major floods. ***a****) A reach of the du Moulin River, Saguenay area, Quebec, several days after the July 1996 flood (flow is from top to bottom). Photograph by G.R. Brooks, taken on 26 July 1996. GSC 1997-42T* ***b****) A reach of the Red River in southern Manitoba following the major 1997 flood (flow is from right to left). Photograph by G.R. Brooks, taken 14 June 1997. GSC 2000-114 In both examples, the river channel is essentially unchanged from the pre-flood morphology. In the du Moulin River example, there is no obvious indication that a major flood occurred only days before the photograph was taken.*

the re-establishment and growth of vegetation. The amount of widening that occurs during a subsequent extreme discharge will depend on how long it has been and how much 'recovery' has occurred since the previous flood event; the widening likely will be minor if there has been little recovery, but possibly major if there has been significant recovery.

Erosion along an alluvial channel by floodwaters can result in the transport of large quantities of sediment. This sediment is carried downstream and deposited in mid-channel bars, point bars, and channel marginal deposits, and on the floodplain as overbank sediments, depending on the local river morphology. Aggradation of the valley bottom may occur along some reaches if flow energy drops because of a marked widening of the channel or valley bottom and/or a significant decrease in channel gradient (Fig. 19). In extreme circumstances, the post-flood river flows entirely on a surface of freshly accreted alluvium (Fig. 19). Aggradation also occurs at a river mouth where the flow enters a lake or the sea and at the base of a steep stream where the gradient drops as it emerges onto an alluvial fan.

Flood damage

Flooding causes loss of life and damages property and infrastructure (Fig. 20). Buildings and infrastructure (e.g. bridges, pipelines) can be damaged structurally or destroyed by fast-flowing water and/or impacts from debris (ice, trees) carried by the current. In extreme circumstances, buildings and bridges may be washed off their foundations and carried downstream. Lateral bank erosion can damage or destroy buildings and infrastructure by undermining them, even when they are situated above the level of inundation (Fig. 21). Bridge abutments or pier supports may be scoured and undermined in areas where they constrict and accelerate the flow. Bridges can also partly dam flow and be overtopped by water, causing the approaches to be washed out (Fig. 22). Floodwaters can wash out roads, highways, and railway lines. Artificial dams may be breached by overtopping flood flows (Fig. 23). In cases where breaching causes the rapid drainage of a reservoir, the addition of escaping reservoir water to the flood runoff can produce a flow far larger than that generated

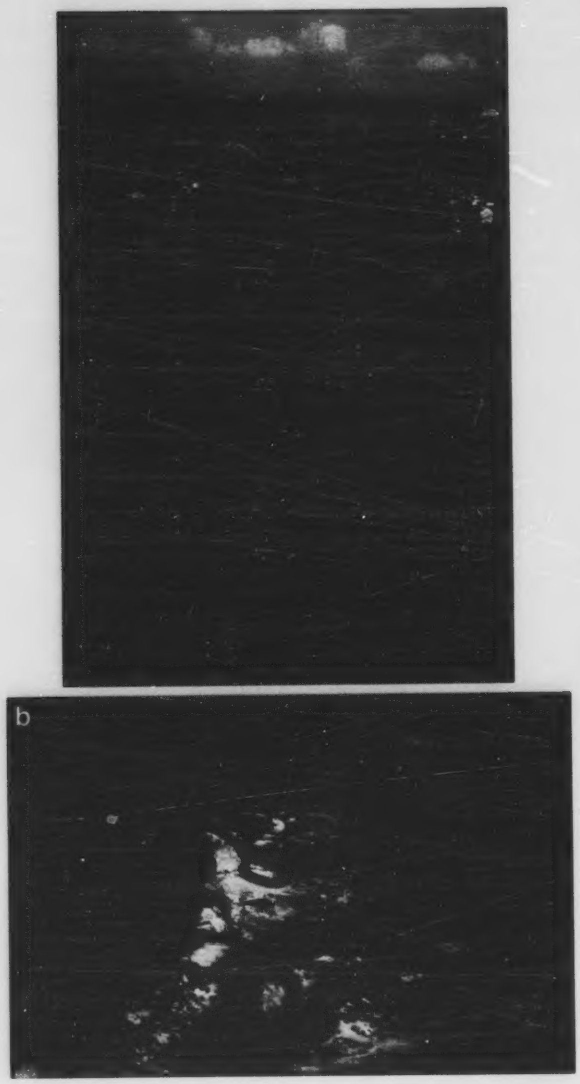


Figure 18. Examples of valley bottoms that underwent large-scale erosion by flood waters. a) The broad post-flood channel of the à Mars River, Saguenay area, Quebec, following the July 1996 flood; view downvalley from about 4 km above the river mouth. Extensive channel widening occurred during the flood through large-scale reworking of the floodplain and lateral bank erosion of terrace deposits. The post-flood channel exhibits a multichannel planform in contrast to the predominantly single-channelled meander planform of the pre-flood river. The morphology of this river reach was completely transformed by the flood. Photograph by G.R Brooks taken on 27 July 1996. GSC 1997-42X b) Effects of the 1983 outburst flood in the Nostetuko River valley, British Columbia. Sediments and trees were stripped from the valley floor by the floodwaters, which were about 5 m deep here. Photograph by J.J. Clague, taken in August 1984. GSC 2000-120

by strictly natural hydrometeorological processes (similar to an outburst flood from the failure of a natural dam). Such floods can have catastrophic effects and impacts on downstream river reaches and communities.

In addition to causing flooding, jammed and floating ice during breakup can be very destructive along rivers prone to ice jamming. Ice may pile up along the channel margins and be pushed along and thrust onto the banks or floodplain to heights well above the water level (Fig. 5). A shearing and thrusting ice mass is a threat to structures adjacent to or spanning a river. Bridges can be damaged or destroyed by ice jams that form behind them.

Significant damage from flooding also can arise even where flood waters are flowing relatively slowly or are stagnant. Inundated buildings (except specially designed 'flood-proof' buildings) and their contents will have water

Figure 19. Extensive aggradation along a broad, relatively gently sloping reach of the Ha! Ha! River, Saguenay area, Quebec, resulting from the July 1996 flood. Several properties are covered by up to 2 m of sand derived from severe upstream channel incision. Photograph by G.R. Brooks. GSC 1997-42KK

Figure 20. Buildings flooded at St. Norbert, Manitoba, during the 1997 Red River flood. Photograph by G.R. Brooks. GSC 2000-115

damage from flooding with the potential for damage increasing with greater water depth (Fig. 20). Any contaminants within the water (e.g. oil, gasoline, raw sewage, chemicals) also will enter buildings and coat floors, walls, and building contents, compounding the water damage. Electrical short circuits may start fires in partly flooded buildings. Sewer backups can flood the basements of dry buildings when the capacity of storm-sewer drainage systems is exceeded by storm rainfall runoff.

Water-damaged buildings require thorough cleanup, decontamination, and restoration following a flood. Allergenic and toxigenic fungi can grow within wet drywall, insulation, books, carpets, and air ventilation systems, and can cause significant health problems to occupants long after the flood. In some cases, it may be best to write off the structure and rebuild on higher ground or in a less flood-prone area. Inundation of agricultural lands in the spring may delay planting, thereby shortening the growing season and affecting the type of crops that can be grown. Planted crops (e.g. winter wheat) can be damaged by persistent inundation, thus affecting crop yields. The harvest of forage crops may be delayed by wet ground conditions after summer floods.

Sediment deposited by flood waters can damage property. Clay and silt carried in suspension will settle out on the floodplain, in low-lying areas, and within inundated buildings, forming a veneer of mud. In some floods, floodplain

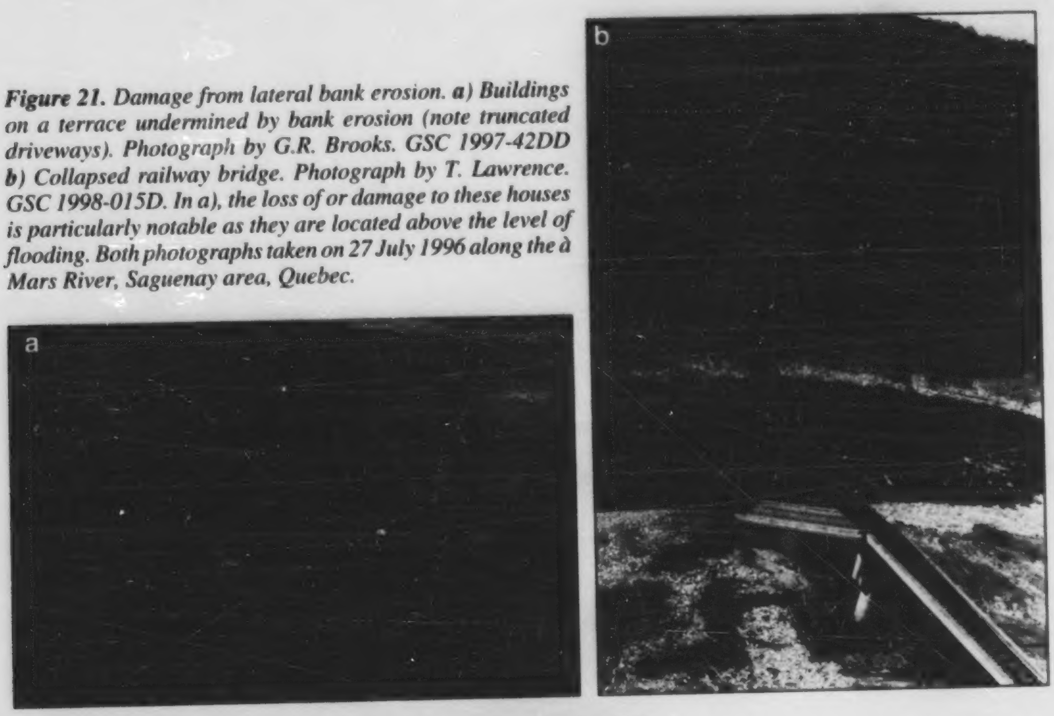


Figure 21. *Damage from lateral bank erosion.* ***a)*** *Buildings on a terrace undermined by bank erosion (note truncated driveways). Photograph by G.R. Brooks. GSC 1997-42DD* ***b)*** *Collapsed railway bridge. Photograph by T. Lawrence. GSC 1998-015D. In a), the loss of or damage to these houses is particularly notable as they are located above the level of flooding. Both photographs taken on 27 July 1996 along the à Mars River, Saguenay area, Quebec.*

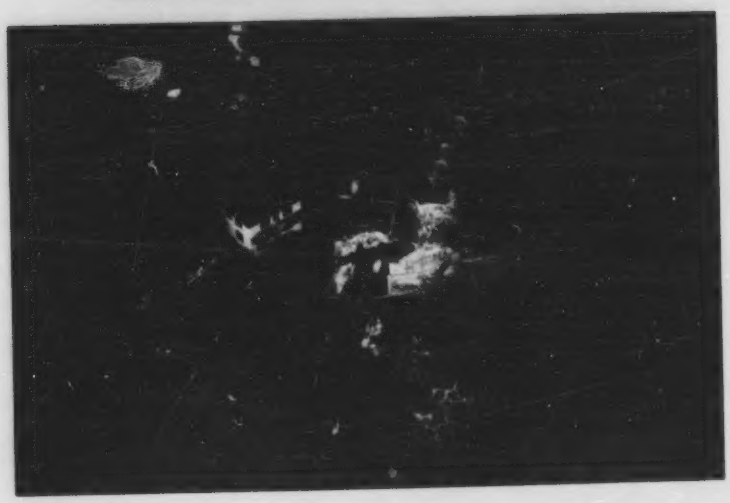


Figure 22.

Washed-out road and two severely damaged houses along the du Moulin River, Saguenay area, Quebec. This damage occurred because the bridge abutments constrict a section of the river causing floodwaters to spill over the bridge. Photograph by G.R. Brooks, taken on 27 July 1996. GSC 1997-42WW

Figure 23.

Post-flood view of the Chute-Garneau dam site along the Chicoutimi River, Saguenay area, Quebec, which was severely damaged during the July 1996 flood. Water spilled over and eroded into unconsolidated marine deposits (silt and clay) forming the right abutment. A new channel carries the flow beside the dam, completely bypassing the penstocks and sluice gates. The dam is left intact, but nonfunctional. Photograph by G.R. Brooks, taken on 26 July 1996. GSC 1997-42N

areas may be buried under sand and gravel. The thickness of sediment deposited on floodplains varies considerably depending on the energy of flow and the amount and calibre of sediment introduced into the river by erosion upstream. In extreme cases, broad sheets of sediment up to several metres thick can be deposited (Fig. 19). Deposition of silt and clay on floodplains can increase soil fertility, but may be detrimental to yields in the short term if crops or pastures are buried.

FLOOD FORECASTING AND FREQUENCY ANALYSIS

Flood forecasting

In general, the specific date and magnitude of hydrometeorological floods cannot be predicted over long periods. For some watersheds, spring discharge levels can be estimated several days in advance using models that incorporate snowpack levels, stream and lake levels, ground moisture content, temperature, wind, evaporation, and weather forecasts. 'Outlooks' of peak spring discharge can be made weeks in advance using similar models with various scenarios of melt rates and rainfall based on climate records. Potential peak flow levels are expressed as probabilities of occurrence. With such estimates, the actual level of peak runoff is dependent on uncertain climatic events that are yet to occur, but worst-case scenarios can be identified and possibly planned for well in advance of the actual flooding.

Flood advisories or warnings can be issued in expectation of severe weather conditions (convective storms, mid-latitude and tropical cyclones). The advisories or warnings can be regional or local, but in most instances are issued only hours in advance of the storm. However, they can provide sufficient time for the initiation of flood-control measures (e.g. the construction of earth and/or sandbag dykes) and, in extreme cases, the evacuation of people, livestock, or property from flood-prone areas.

It is currently impossible to predict the occurrence of an ice jam and associated flood along a given river. However, data from previous years can be used to relate the time and magnitude of the peak water level during breakup to the current hydrometeorological conditions. Such data are site-specific because some locations along a given river are more prone to ice jamming than others. However, once an ice jam has formed somewhere along a river, empirical methods can be used to predict the rise in water levels.

Forecasting floods caused by the failure of natural dams is a problem of timing rather than magnitude. A number of empirical formulae exist for the different types of natural dams for estimating peak discharge of outburst floods. They express maximum discharge as a function of the potential energy or volume of the impounded water. In addition, outburst floods may be predicted using dam-break models, provided that the initial water volume and the topography of the flood path are known.

When a landslide dam forms, the extent of the resulting impoundment may be readily estimated from the elevation of the debris crest. Warning may be given to inhabitants downstream from an impending outburst flood, should one be expected. An estimate of the time of filling can be made from the volume of the impoundment and the rate of water inflow. Moraine dam bursts are highly unpredictable, although they generally occur in the summer following a period of warm weather. Jökulhlaups from glacier dams generally occur in the summer or early fall.

Flood frequency analysis

For planning purposes, the occurrence of storm-rainfall, rain-on-snow, and snowmelt floods along a given river is treated as a probability problem using existing records of annual peak discharge. The recurrence interval (or return period) of each annual maximum flow in the discharge record is calculated and the discharge-recurrence interval for the entire data set is plotted (Fig. 24). The recurrence interval for each given magnitude of flow represents the average time of occurrence between flows of equal or larger magnitude.

Estimates of the recurrence interval of low-frequency, high-magnitude floods exceeding those previously recorded are obtained by extrapolating a curve fitted to the data. The flood frequency-magnitude distribution reflects the existing discharge record and will be modified by future floods, particularly ones exceeding those previously recorded. From the distribution, the magnitude of a flood with a given probability of occurrence can be estimated for floodplain management or engineering design purposes (e.g. the '100-year flood' — the magnitude of flow corresponding to a recurrence interval (or return period) of an average of once in 100 years). However, such flood frequencies are statistical probabilities, and recurrence intervals are not predictions of when specific discharges will actually occur. There is always a chance that a major flood will occur in any given year and will be larger than, for example, the 50-year flood, the 100-year flood, or the 500-year flood. Whereas flood recurrence estimates provide a reasonable approximation of the 'expected' return period of a flood of a given or larger magnitude, there are complicating factors, as summarized in Table 5.

FLOOD MITIGATION

The simplest and least expensive way to reduce the damaging effects of flooding is to identify and carefully manage flood-prone lands. Since 1975, under the federal-provincial Flood Damage Reduction Program, floodplain mapping has been undertaken along many river reaches prone to damaging

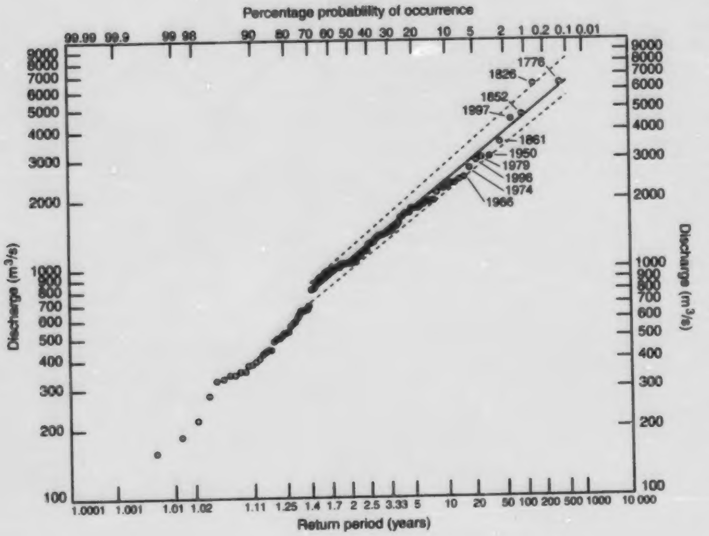


Figure 24.

Frequency-magnitude plot of the annual peak flows of the Red River, Manitoba, for the years 1776, 1826, 1851, 1861, 1875 to 1878, 1880 to 1885, and 1892 to 1997 (discharge data for the Red River at Redwood Bridge station; Manitoba Department of Natural Resources). A best-fit line and 95% confidence limits are fitted to the data. The ten highest flows in the record are labelled by year. The discharges for 1776, 1826, 1852, and 1861 are estimates based on historic accounts of floods. The discharges in the other years are measured flows, but post-1969 they have been corrected to take into account flow diversions (Red River floodway, Portage diversion) and storage in an artificial reservoir (Shellmouth dam). The flood in May 1997 that caused widespread flood damage in the Red River valley is the fourth highest in the record and has a recurrence interval of 62 years on the basis of this data set.

Table 5. Factors complicating estimates of flood occurrence interval or return periods.

1	The record of annual peak discharge may be the product of several mechanisms, i.e. it is a mixed rather than a single population of floods. In such circumstances, the resulting plot of discharge-flood recurrence interval may be nonlinear and/or have considerable scatter. Extrapolation based on a regression may over- or underestimate the recurrence of extreme events.
2	The discharge record is short and does not include a 'large' event. Estimates of the recurrence interval of high-magnitude discharges based on such a record may be significantly lower than from a record of longer duration that incorporates one or more 'large' events.
3	The occurrence of an anomalously large flood caused by an unusual event (e.g. a tropical storm well off the normal track) may appear to have a recurrence interval on the order of thousands of years on the basis of existing discharge-flood recurrence plots. The estimated recurrence intervals of such floods, however, are unreliable and should be viewed with caution. The flood mechanism may not be adequately represented in the known flood record and floods of this size may occur much more frequently than the data suggest.
4	Ice jams can elevate water levels well above those produced solely by the discharge at the time. Ice-jam floods are usually examined from the perspective of historical stage levels, if data are available. Such data are site specific because of the influence of local factors in the formation of ice jams.
5	Climate change can alter the temporal pattern of flood recurrence so that floods of a given magnitude may occur at a different frequency in the future than in the past.

floods, based on a discharge with a 100-year recurrence interval. Other discharge criteria are used locally; for example, British Columbia generally uses a 200-year recurrence interval, whereas some parts of Ontario and Manitoba use the discharge of a past regional flood. Land use and building construction are carefully regulated in areas susceptible to flooding. In this context, it is common to see parks, playing fields, and natural areas situated on floodplains in urban areas. These types of land use are compatible with periodic flooding with minor consequences.

Although avoiding flood-prone areas seems straightforward, considerable development has occurred on floodplains in Canada for a number of interrelated reasons, including the following: local physiography and historic economic factors promoted the concentration of development in river valleys; there was insufficient awareness of local flood hazards during initial settlement; development preceded the formal designation of floodplain areas; there is a continued willingness to assume the risk of occupying flood-prone lands; and existing floodplain regulations and building standards are not always enforced. Once initiated, development tends to attract further development, compounding potential flood damage. In many cases, it is not economically or politically feasible to relocate the development, and engineering options are initiated to address the flooding problem. Common engineered flood-control measures are summarized in Table 6 and depicted in Figure 25.

Flood-control structures are effective within their design parameters, as long as they are adequately maintained and there are no changes in the drainage basin upstream that increase the flow of water through the watershed (thus increasing the peak discharge downstream), or changes downstream that accentuate the flow stage (or depth) of a given discharge at the flood-control structures. Such structures, however, are not without impact. They can accentuate flooding along reaches adjacent to the protected reach by raising flood levels upstream and/or by accelerating the movement of water downstream.

Many Canadian towns and cities that encroach on flood-prone lands and have had flooding problems in the past are protected by flood-control structures and are generally unaffected by moderate to large floods. The structures, however, can lead to a sense of complacency about the local flood hazard and often promote further development on the formerly flood-prone lands. There is always the possibility that an extreme flood exceeding the 'design' discharge (e.g. the 100-year flood) will occur, overwhelming the protective structures and inundating the protected land. Such a disaster is virtually inevitable over the long term because it is not economically feasible to construct flood-control structures that can contain the maximum possible discharge of a given river.

Several measures can mitigate the formation and severity of ice jams. Before breakup, an intact ice surface can be dusted with a thin layer of dark powder (sand, fly ash, or another environmentally safe substance) to increase the absorption of solar radiation and thereby promote in situ melting and weakening of the ice cover. Ice blasting and ice cutting can be used individually or in combination to break up an intact ice cover. The goal of both methods is to prevent jamming at a particular location by breaking intact ice into pieces to promote transport downstream or to remove a solid ice cover behind which a jam could form. On larger rivers, icebreakers or air-cushion vehicles (hovercrafts) can be used to break up an ice cover. At locations vulnerable to major ice-jam flooding or ice damage, or where ice accumulation greatly interferes with hydroelectric power generation or navigation, structural measures can be implemented, including the use of pier-mounted booms, groins, and jetties that

Table 6. Common flood-control structures (*see* Figure 25).

Platforms	Artificial mounds of earth upon which buildings are constructed. Platforms provide local flood protection by physically raising the ground surface to a height at or above the designated 'flood' level.
Dykes	Permanent or temporary linear embankments built along channel courses or in front of low-lying areas. Dykes provide a physical barrier that keeps water either confined within channels or out of low-lying areas. When built along channels, dykes increase the channel cross-sectional area and thus the channel discharge capacity. Ring-dykes provide local protection from flooding to areas ranging in scale from a single house to a small town. Dykes are the most common flood-control structure in Canada.
Channel improvements	A variety of measures including deepening and widening the channel, changing the channel shape and roughness, cementing the channel perimeter, and straightening the channel. All such improvements are intended to increase the efficiency and discharge capacity of the channel, allowing it to convey more water.
Diversions	Flow is fully or partly diverted along an artificial channel that either bypasses an area, rejoining the river downstream, or enters a different watershed. In both cases, the diversion of flow is regulated by a control dam that may be supplemented by a dyking system. The intent of a diversion is to provide an alternative route for excess discharge away from the protected low-lying area.
Flow regulation	Water is temporarily stored within artificial reservoirs and/or retention basins. This storage reduces the magnitude of snowmelt and storm-rainfall flowing downstream. The subsequent gradual release of stored water increases discharge during low-flow periods. Flow regulation is most effective during the spring when reservoirs are drawn down to maximize storage capacity in anticipation of the impending snowmelt runoff.

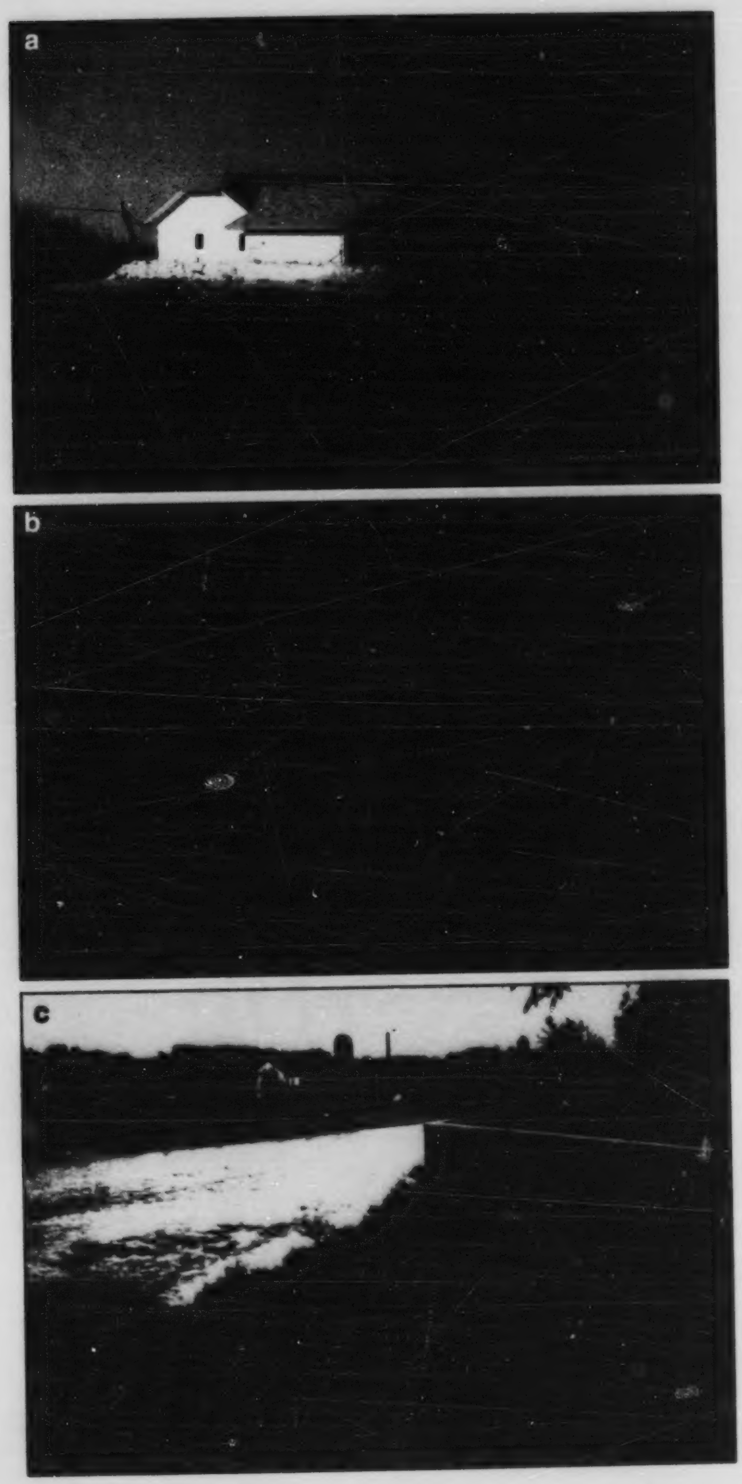


Figure 25.

Examples of flood-control measures: ***a****) building situated on a earth platform, further protected by a sandbag ring-dyke, and surrounded by floodwater (Grande Pointe, Red River valley, Manitoba, 1997). Photograph by G.R. Brooks. GSC 2000-060;* ***b****) ring-dykes around a town that remains dry despite widespread inundation of the surrounding area (Rosenort, Red River valley, Manitoba, 1997). Photograph by G.R. Brooks. GSC 2000-116;* ***c****) river channel that has been artificially deepened and banks armoured with concrete and dyked to increase the discharge capacity through an urban area (Grand River, Cambridge, Ontario; photograph courtesy of the Grand River Conservation Authority);* ***d****) flood-control dam and reservoir (Kelso Conservation Area, Ontario). Photograph by G.R. Brooks. GSC 2000-117; and* ***e****) floodway diverting flow from a river channel to reduce the discharge downstream (Red River Floodway, Manitoba, during the 1997 flood). Photograph by G.R. Brooks. GSC 2000-118* constrict a channel, artificial islands, ice-control dams and weirs, and ice booms. All are intended to stabilize an ice cover or prevent the movement of ice downstream, thus promoting in situ melting. Structural measures are generally reliable, but costly, because they must be carefully designed and maintained.

The prevention of flooding by lakes impounded by natural dams has been attempted on only a few occasions in Canada. Spillways have been excavated across several landslide dams to control overtopping and prevent rapid erosion. Trenches were cut across the Bear Glacier, which formerly dammed Strohn Lake, near Steward, British Columbia, in an effort to drain the lake; these measures, however, were unsuccessful in preventing jökulhlaups. In contrast, annual pumping of a small supraglacial lake at Cathedral Mountain, near Kicking Horse Pass, British Columbia, has eliminated jökulhlaups along Cathedral gulch. The jökulhlaups had triggered debris flows that travelled several kilometres to the bottom of Kicking Horse River valley, blocking traffic along the Trans-Canada Highway and Canadian Pacific Railway main line.

CLIMATE CHANGE

Climate determines the range of flows of a river (excepting outburst floods). Climate, however, is not constant. Since the 1860s, there has been a net increase in global temperatures. It is now recognized that some of this warming is the result of burning fossil fuels, which is increasing the amount of greenhouse gases in the atmosphere. Global climate models indicate that mean global temperature may increase several degrees by the middle of the twenty-first century, with shifts in the amount and distribution of precipitation.

Climate change has a number of implications for river flooding. Most importantly, the frequency of extreme flood events may increase, with obvious implications for flood mitigation. The mapping of floodplains and the design of engineered structures are based on a particular flood magnitude (the design flood) with a given recurrence interval (e.g. the 100-year flood; *see* 'Flood mitigation'). An increase in the frequency of extreme floods lowers the recurrence intervals of large floods and alters the original frequency-magnitude relationship used to establish the design flood. For example, the recurrence interval of a magnitude 200-year flow may be lowered to 100 years, or the original 100-year flow may be reduced to a recurrence interval of 75 years. This means that there is an increased chance of a future extreme flood inundating low-lying areas outside the margin of a designated floodplain or overtopping a flood-control structure. Thus, climate change is expected to increase the risk of occupying flood-prone lands along many rivers in the future.

With respect to natural dams, the most sensitive to climate change are those related to glaciers, i.e. glacier and moraine dams. Late nineteenth- and twentieth-century climate warming has resulted in a dramatic loss of glacier ice in the Canadian Cordillera. As noted above, the formation and failure of moraine dams are directly related to this aspect of climate change. The relationship between glacier retreat and the formation and failure of glacier dams is complex. Initial glacier retreat leads to the formation of glacier-dammed lakes; further retreat leads to jökulhlaups, but still further retreat and thinning may create conditions under which the lake ceases to exist or is of such limited size that a jökulhlaup cannot occur.

ACKNOWLEDGMENTS

We thank Chris Tucker (Emegency Preparedness Canada) for permission to use the EPC disaster database, and John Newton for information about flooding in arctic communities. Comments on the manuscript by Joe Desloges, Raimo Kallio, and Ted Lawrence are sincerely appreciated.

REFERENCES

Clague, J.J. and Evans, S.G.
1994: Formation and failure of natural dams in the Canadian Cordillera; Geological Survey of Canada, Bulletin 464, 35 p.
Environment Canada
1988: Flood events in Canada 1983–1987; Water Planning and Management Branch, Inland Waters Directorate, Environment Canada, 84 p.
Kindervater, A.D.
1980: Flooding events in Newfoundland and Labrador a historical perspective; Water Planning and Management Branch, Inland Waters Directorate, Atlantic Region, Environment Canada, 80-WPMB-4, 142 p.
1985: Flooding events in New Brunswick a historical perspective; Water Planning and Management Branch, Inland Waters Directorate, Atlantic Region, Environment Canada, IWD-AR-WPMB-84-65, 247 p.
1988: Flooding events in Nova Scotia a historical perspective; Inland Waters Directorate, Atlantic Region, Environment Canada, 147 p.
Mathews, W.H. and Clague, J.J.
1993: The record of jökulhlaups from Summit Lake, northwestern British Columbia. Canadian Journal of Earth Sciences, v. 30, p. 499–508.
Saskatchewan Water Resources Commission
1972: A historical review of flooding in the Qu'appelle River basin 1852–1971; Qu'appelle Flood Study Appendix C; Investigation and Planning Branch, Saskatchewan Water Resources Commission, 107 p.
Septer, D. and Schwab, J.W.
1995: Rainstorm and flood damage, northwestern British Columbia, 1891–1991; Research Branch, British Columbia Ministry of Forests, 196 p.
Tudor, C.
1997: EPC electronic disaster database and its characteristics, Project 97; Office of the Senior Scientific Advisor, Emergency Preparedness Canada, 19 p.

SUGGESTED READING

General reviews

Andrews, J.
1993: Flooding — Canada water book; Environment Canada, Ottawa, Ontario, 171 p.
Church, M.
1988: Floods in cold climates; *in* Flood Geomorphology, (ed.) V.R. Baker, R.C. Kochel, and P.C. Patton; J. Wiley and Sons, New York, p. 205–229.
Matthai, H.F.
1990: Floods; *in* Surface Water Hydrology, (ed.) M.G. Wolman and H.C. Riggs; Geological Society of America, The Geology of North America, v. O-1, Boulder, Colorado, p. 97–120.
Watt, W.E.
1989: Hydrology of floods in Canada: a guide to planning and design; Associate Committee on Hydrology, National Research Council, 245 p.

Impact of climate change

Ashmore, P. and Church, M.
in press: The impact of climate change on rivers and river processes in Canada; Geological Survey of Canada, Bulletin 555.

Snowmelt-runoff and storm-rainfall floods

Baker, V.R. and Costa, J.E.
1987: Flood power; *in* Catastrophic flooding, (ed.) L. Mayer and D. Nash; Allen and Unwin, London, p. 1–21.
Costa, J.E. and O'Connor, J.E.
1995: Geomorphically effective floods; *in* Natural and Anthropogenic Influences in Fluvial Geomorphology, (ed.) J.E. Costa, K.W. Potter, and P.R. Wilcock; American Geophysical Union, Geophysical Monograph 89, p. 45–56.
Kochel, R.C.
1988: Geomorphic impact of large floods: reviews and new perspectives on magnitude and frequency; *in* Flood Geomorphology, (ed.) V.R. Baker, R.C. Kochel, and P.C. Patton; J. Wiley and Sons, New York, p. 169–187.
National Oceanic sand Atmospheric Administration
1994: The great flood of 1993; Natural Disaster Survey Report, National Oceanic and Atmospheric Administration, various pages.

Ice-jam floods

Beltaos, S. (ed.)
1995: River ice jams; Water Resources Publications, Highlands Ranch, Colorado, 372 p.
Davar, K.S, Beltaos, S., and Pratte, B.
1996: Primer on hydraulics of ice covered rivers; Canadian Committee on River Ice Processes (CGU-HS) and Environment Canada, Fredericton, New Brunswick, 445 p.
Gerard, R.
1990: Hydrology of floating ice; *in* Northern Hydrology: Canadian Perspectives, (ed.) T.D. Prowse and C.S.L. Ommanney; National Hydrology Research Institute Science Report no. 1, Environment Canada, p. 103–134.
MacKay, D.K. and Mackay, J.R.
1973: Location of spring ice jamming on the Mackenzie River, N.W.T.; *in* Hydrologic Aspects of Northern Pipeline Development; Environmental-Social Committee, Northern Pipelines, Task Force on Northern Oil Development, Report 73-3, p. 233–257.
Michel, B.
1978: Ice Mechanics; Les Presses de l'Université Laval, Québec, Quebec, 499 p.
New Brunswick Subcommittee on River Ice
1989: New Brunswick river ice manual; Inland Waters Directorate, Environment Canada, New Brunswick Department of the Environment, 18 p.
Prowse, T.D.
1984: Liard and Mackenzie river ice break-up, Fort Simpson region, N.W.T., 1983; Northern Hydrology Section, Surface Water Division, National Hydrology Research Institute, Environment Canada, 72 p.

Floods caused by natural dams

Blown, I. and Church, M.
1985: Catastrophic lake drainage within the Homathko River basin, British Columbia; Canadian Geotechnical Journal, v. 22, p. 551–563.
Clague, J.J. and Evans, S.G.
1997: The 1994 jökulhlaup at Farrow Creek, British Columbia, Canada; Geomorphology, v. 19, p. 77–87.
Clague, J.J. and Rampton, V.N.
1982: Neoglacial Lake Alsek; Canadian Journal of Earth Sciences, v. 19, p. 94–117.

Clague, J.J., Evans, S.G., and Blown, I.
1985: A debris flow triggered by the breaching of a moraine-dammed lake, Klattasine Creek, British Columbia; Canadian Journal of Earth Sciences, v. 22, p. 1492–1502.

Costa, J.E. and Schuster, R.L.
1988: The formation and failure of natural dams; Geological Society of America Bulletin, v. 100, p. 1054–1068.

Desloges, J.R. and Church, M.
1992: Geomorphic implications of glacier outburst flooding: Noeick River valley, British Columbia; Canadian Journal of Earth Sciences, v. 29, p. 551–564.

Evans, S.G.
1986: Landslide damming in the Cordillera of Western Canada; *in* Landslide Dams: Processes, Risk, and Mitigation; (ed.) R.L. Schuster; American Society of Civil Engineers, Geotechnical Special Publication, no. 3, p. 111–130.
1987: The breaching of moraine-dammed lakes in the southern Canadian Cordillera; *in* Proceedings, International Symposium on Engineering Geological Environment in Mountainous Areas, v. 2, p. 141–150.

Evans, S.G. and Brooks, G.R.
1994: An earthflow in sensitive Champlain Sea sediments at Lemieux, Ontario, June 20 1993, and its impact on the South Nation River; Canadian Geotechnical Journal, v. 31, p. 384–394.

Heginbottom, J.A.
1984: The bursting of snow dam, Tingmisut Lake, Melville Island, Northwest Territories; *in* Current Research, Part B; Geological Survey of Canada, Paper 84-1B, p. 187–192.

Jackson, L.E.
1979: A catastrophic glacial outburst flood (jökulhlaup) mechanism for debris flow generation at the Spiral Tunnels, Kicking Horse River basin, British Columbia; Canadian Geotechnical Journal, v. 16, p. 806–813.

Mathews, W.H.
1965: Two self-dumping ice-dammed lakes in British Columbia; Geographical Review, v. 55, p. 46–52.
1973: Record of two jökullhlaups [*sic*]; *in* Symposium on the Hydrology of Glaciers, (ed.) J.W. Glen, R.J. Adie, and D.M. Johnson; International Association of Scientific Hydrology, Publication 95, p. 99–110.

Walder, J.S. and Costa, J.E.
1996: Outburst floods from glacier-dammed lakes: the effect of mode of lake drainage on flood magnitude; Earth Surface Processes and Landforms, v. 21, p. 701–723.

Other publications of interest

Canada Mortgage and Housing Corporation
1994: Cleaning up your house after a flood; Canada Mortgage and Housing Corporation, Canada, 39 p.

APPENDIX A

Significant flood disasters known to have occurred in Canada during the twentieth century (data primarily from Emergency Preparedness Canada with modifications)

Date (y/m/d)	Location	Damage ($M)*	Deaths	Comment
1902/03/17	Saint John River basin, New Brunswick	0.025	1	Flood; Saint John River basin, New Brunswick, 17–23 March 1902; an unusually early breakup in March and downstream movement of ice cover on the Saint John River due to heavy rains and warm temperatures raised water levels, leading to 15 ice jams; extensive damage including washouts of bridges, roads, and railways; one dam destroyed; cattle drowned; the lumber industry was also affected — damage to mills and loss of stock were reported; province-wide damage estimated at $25 000. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1904/04	Qu'appelle River basin, Saskatchewan	unknown	0	Flood; Qu'appelle River basin, Saskatchewan, April 1904; rapid and unusually large runoff following above-normal fall and winter precipitation; flooding in Regina, Lumsden-Crave area, Moose Jaw; houses and farms flooded, bridges washed out, dam destroyed, railway line inundated (Saskatchewan Water Resources Commission, 1972)
1904/04/24	Red River, Manitoba	unknown	0	Flood; Winnipeg, Manitoba; severe flooding in the Winnipeg area of the Red River valley. (Emergency Preparedness Canada database, December 1997)
1909/01/04	Chester, New Brunswick	0.1	1	Flood; New Brunswick, 4–7 January 1909; mild weather and heavy rains caused ice to travel and possibly jam, creating floods that caused expensive damage to bridges and roads; one dam damaged; a nine-year-old child fell from a small bridge and drowned in a flood-swollen stream near Chester; $100 000 damage. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1912/04/01	southern Ontario	0.5	0	Flood; southern Ontario, 1–8 April 1912; widespread flooding caused minor damage and flooded several roads and railways; river dam systems were damaged and many utility services were disrupted; businesses in several communities were forced to close during the flooding; total reported damages reached $0.5 million. (Emergency Preparedness Canada database, December 1997)
1913/05/02	Assiniboine River, Manitoba	unknown	0	Flood; Assiniboine River, Manitoba; large-scale flooding. (Emergency Preparedness Canada database, December 1997)
1915	Alberta–Saskatchewan border on North Saskatchewan River	unknown	0	Flood; North Saskatchewan River, Alberta and Saskatchewan; the combination of rain in the foothills and snowmelt resulted in the highest recorded spring flows on the North Saskatchewan River; the cities of Edmonton, Alberta, and Prince Albert, Saskatchewan, were flooded. (Emergency Preparedness Canada database, December 1997)
1915/07/08	northeastern New Brunswick	0.25	0	Flood; northeastern New Brunswick; flooding was reported on 8–9 July 1915; two bridges washed out on the Black River; many logs swept out to sea on the Miramichi River; damage estimated at $250 000 (1915 dollars). (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1916/04/24	Red River, Manitoba	unknown	0	Flood; Winnipeg, Manitoba; severe flooding in the Winnipeg area of the Red River valley. (Emergency Preparedness Canada database, December 1997)
1916/04/25	central Ontario	1	0	Flood; central Ontario; flooding caused direct damages totalling $1 million to railways, dams, and businesses. (Emergency Preparedness Canada database, December 1997)
1917/11/28	northwestern British Columbia	unknown	0	Flood; northwestern British Columbia, 28 October–19 November 1917; caused by wet and stormy weather; widespread flooding along the Skeena and Nass rivers; numerous bridges damaged; village of Aiyanish 'wiped out', cattle and horses drowned, crops washed away, and many houses and provisions washed away. (Septer and Schwab, 1995)

*All cost of damages relate to the year of the disaster and are not corrected for inflation.

Date (y/m/d)	Location	Damage ($M)*	Deaths	Comment
1920/03/17	southeastern Ontario	0.9	0	Flood; southeastern Ontario, 17–21 March 1920; widespread flooding caused minor damage and flooded roads and railways; direct damage costs reached $0.9 million. (Emergency Preparedness Canada database, December 1997)
1922	Winnipeg, Manitoba	unknown	0	Flood; Winnipeg, Manitoba; severe flooding in the Winnipeg area along the Assiniboine River. (Emergency Preparedness Canada database, December 1997)
1922/06/17	Saint John River basin, New Brunswick	0.5	0	Flood; Saint John River basin and northeastern river basins, New Brunswick, 17–25 June 1922; flooding reported following six days of heavy rainfall; damage confined to northern part of province; two dams and many bridges destroyed; roads, bridges, and railways damaged; houses swept away and crops damaged; damage costs estimated at $500 000. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1923/04	Winnipeg, Manitoba	unknown	0	Flood; Winnipeg, Manitoba, April–May; severe flooding in the Winnipeg area along the Assiniboine River. (Emergency Preparedness Canada database, December 1997)
1923/04/29	Saint John River basin, New Brunswick	3.75	2	Flood; New Brunswick; snowmelt, ice jams, and heavy rainfall caused flooding in New Brunswick from 29 April to 9 May 1929; flooding occurred in the Saint John River basin as well as in river basins in the northeast and southwest; the situation worsened with the creation of log jams from the lumber industry along the rivers systems; roads and railway damaged; 57 bridges damaged or destroyed; lumber mills flooded and logs swept away; damage estimated at $3.75 million; two drownings. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1925/05/07	Dawson, Yukon Territory	unknown	0	Flood; Dawson, Yukon Territory; ice broke on the Klondike River on May 7 and on the Yukon River on 9 May; an ice jam formed, causing severe flooding (4–5 ft. or 1.2–1.5 m high) in Dawson. (Emergency Preparedness Canada database, December 1997)
1928	Chaudière River, Quebec	unknown	3	Flood; Rideau, Chaudière, and Quyon rivers, Quebec; several people were killed by the flooding when these rivers overflowed their banks. (Emergency Preparedness Canada database, December 1997)
1929/01/18	southern Ontario	0.25	0	Flood; southern Ontario, 18–20 January 1929; widespread flooding caused damage and flooded roads and railways; flooding forced several businesses to close; over $250 000 in direct damage was reported. (Emergency Preparedness Canada database, December 1997)
1929/04/05	southern Ontario	1.3	0	Flood; southern Ontario, 5–9 April 1929; widespread flooding caused damage and flooded roads and railways; flooding forced several businesses to close in communities along the Grand and Saugeen rivers and in Toronto area. (Emergency Preparedness Canada database, December 1997)
1933/05/04	New Brunswick and eastern Quebec	unknown	3	Flood; snowmelt, heavy rains, and possibly ice jams contributed to flooding throughout New Brunswick and eastern Quebec, 4–6 May 1933; flood conditions along the Saint John River; houses flooded; three lives lost in Blue River, Tracey Brook (near Kedgwick), and Edmundston. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1933/10/25	southern New Brunswick	unknown	7	Flood; heavy autumn rainfall in southern New Brunswick, 25–26 October 1933; extensive damage to roads, dams (two), and bridges (at least twelve); seven stream drivers who were trapped on a sandbar were swept to their death in Saint John County along the northwest branch of the Big Salmon River. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)*	Deaths	Comment
1934/04/16	New Brunswick	0.232	0	Flood; New Brunswick, 16–24 April 1933; continuous rainfall for several days, snowmelt, and ice jams; extensive flood damage to houses, industry, and infrastructure in the Saint John, Miramachi, and Restigouche river basins; damage costs estimated at $232 000. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1935/01/10	Nova Scotia	1	2	Storm; Nova Scotia, 10–15 January 1935; unusually high temperatures, torrential rains, and ice jams; province hard hit by flooding; dykes and highways damaged; 25 bridges destroyed; houses inundated; roads flooded; 'scores' of farm animals drowned; two people drowned; damages estimated at $1 million. (Kindervater, 1988)
1936/03/10	southern Ontario	0.35	0	Flood; southern Ontario; 10–12 March 1936; widespread flooding caused damages and flooded roads and railways; flooding forced several businesses to close; several dams were also damaged with total costs estimated at more than $350 000. (Emergency Preparedness Canada database, December 1997)
1936/03/16	Saint John River basin, New Brunswick	1.9	0	Flood; New Brunswick; 16–25 March 1936; snowmelt, ice jams, and rainfall caused flooding across New Brunswick and resulted in severe damage; 22 ice jams were reported along the Saint John River basin, some extending 10 km upstream; a major ice jam formed on the Fredericton public wharf and across the Saint John River near the Fredericton Highway bridge; 28 bridges in the province were partly or completely destroyed; roads and railways were inundated; numerous houses, businesses, and barns were flooded or destroyed; lumbermill stock and livestock were lost; province-wide damage estimated at $1.9 million (1936 dollars). (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1936/05/29	northwestern British Columbia	unknown	0	Flood; northwestern British Columbia, 29 May–3 June 1936; widespread flooding in British Columbia, the Yukon Territory, and Alaska caused by a sudden rise in temperature melting snowpack; flooding on the Skeena, Taku, Tusequah, Stikine, and Nass rivers; Usk and Remo along the Skeena River most seriously 'stricken' communities; houses damaged and destroyed at both places and flood damage caused food shortages; flood damage also reported in Hazelton and Telegraph Creek; railway line between Prince Rupert and Smithers damaged by washouts; bridges damaged and destroyed and roads washed out throughout region. (Septer and Schwab, 1995)
1936/09/19	Nova Scotia	0.1	1	Storm; Nova Scotia, 19–20 September 1936; a storm with hurricane force hit, causing flooding throughout province; most low-lying fields in eastern Nova Scotia under water; houses flooded, highways inundated, bridges destroyed, crops damaged, railways washed out; bulk of damage in Shelburne, Yarmouth, Queens, Lunenburg, and Halifax counties; four people drowned — one on land, three at sea; damage estimated at $100 000. (Kindervater, 1988; Emergency Preparedness Canada database, December 1997)
1937/01/13	southern Ontario	unknown	0	Flood; southern Ontario, 13 and 14 January, 1937; widespread flooding caused damage and flooded roads and railways; flooding forced several businesses to close. (Emergency Preparedness Canada database, December 1997)
1937/04/26	London, Ontario	4	0	Flood; southern Ontario, especially on the Thames River in London, 26–28 April 1937; the river flooded, leaving 4000 people homeless and causing over $4 million in direct damage; roads and railways were flooded and businesses were closed. (Emergency Preparedness Canada database, December 1997)
1941/08/02	Burin Peninsula, Newfoundland	unknown	2	Flood; flooding in Burin Peninsula, St. John's, and Little St. Lawrence, Newfoundland, was caused by torrential rains; two employees of United Towns Electric Company were drowned when a dam broke; the total rainfall was reported to be 98.3 mm for a 66 hour period. (Kindervater, 1980; Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1942/09/22	Charlottetown, Prince Edward Island	unknown	0	Storm; Charlottetown, Prince Edward Island; a storm, remnant of a dissipating hurricane, left 163.8 mm of rain and damage in Charlottetown. (Emergency Preparedness Canada database, December 1997)
1947/02/05	Elgin, New Brunswick	unknown	2	Flood; New Brunswick, 5–6 February 1947; rain, mild temperatures, and ice jams caused flooding of large sections of Kings County and surrounding area; two children were swept away and drowned in Sheffar Brook, near Elgin, Albert County. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1947/04/05	southern Ontario	1.2	0	Flood; southern Ontario, 5–12 April 1947; widespread flooding occurred on several river systems (including the Thames, Grand, Rideau, Credit, and Saugeen rivers) as well as in the Lake St. Clair and Lake Nipissing regions; over $1.2 million in direct damage reported; several roads and railways were forced to close due to flooding; utilities and businesses were also disrupted. (Emergency Preparedness Canada database, December 1997)
1947/04/30	Bristol, New Brunswick	unknown	2	Flood; heavy rains, mild temperatures, and snowmelt contributed to heavy runoff and overflow of the Saint John River, New Brunswick, 30 April–10 May 1947; two people were killed in a train wreck near Bristol when the track was undermined by the Saint John River. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1948/03/16	southern Ontario	10	0	Flood; southern Ontario, 16–22 March 1948; communities in the Lake St. Clair region, southeastern Ontario (on the Maitland, Grand, and Thames rivers), south-central Ontario (Toronto area), and southwestern Ontario (Rideau River) had flood damage; transportation systems were shut down and businesses were closed; dams were damaged on the Maitland and Thames rivers and in Toronto; direct damage was estimated at $10 million. (Emergency Preparedness Canada database, December 1997)
1948/05/31	Fraser River lowland, southwestern British Columbia	15	10	Flood; Fraser River, British Columbia; the flood ripped apart dykes; almost 10 per cent of the area of the Fraser River valley (55 000 acres) was flooded; 10 people died and 200 families were left homeless; 3000 buildings were destroyed and 82 bridges were washed out; the total cost of the disaster was estimated at $15 million. (Emergency Preparedness Canada database, December 1997)
1948/09/14	St. John's, Newfoundland	unknown	1	Flood; St. John's, Newfoundland; torrential rains caused at least three houses to collapse when hit by mudslides; in one area, it was necessary to rescue residents by boat, one child died. (Kindervater, 1980; Emergency Preparedness Canada database, December 1997)
1950/03/24	southern Ontario	0.1	0	Flood; southern Ontario, 24–29 March; widespread flooding affected southern Ontario including Toronto; many roads were flooded and businesses were closed along the Grand River, causing an estimated $100 000 in damage. (Emergency Preparedness Canada database, December 1997)
1950/04/04	southern Ontario	unknown	0	Flood; southern Ontario, 4–8 April 1950; widespread flooding across southern Ontario led to the closing of several roads. (Emergency Preparedness Canada database, December 1997)
1950/04/20	Saint John River basin, New Brunswick	0.5	0	Flood; New Brunswick, 20–24 April 1950; heavy rains and spring breakup caused ice jams and consequent flooding of several rivers throughout New Brunswick; damage to bridges, rail lines, and roads estimated at $500 000, with probably over half of that in the Saint John River basin. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)ᵃ	Deaths	Comment
1950/05/05	Red River, Manitoba	100	1	Flood; Winnipeg and Red River valley, Manitoba, 5–25 May 1950; 1 dead, 107 000 evacuated, 5000 buildings damaged, $100 million in damage; a major flood occurred in April, May, and June 1950; melt of a thick snow cover caused the Red River to reach flood levels by 22 April; a heavy rainfall in early May caused the river to stay above flood stage for 51 days; the water reached 4.6 m deep in low-lying areas; 60 000 people were forced to evacuate as one tenth of the city of Winnipeg was covered in water; a volunteer pump operator drowned; damage estimates in 1957 dollars were $125.5 million. (Emergency Preparedness Canada database, December 1997)
1950/08/22	Nova Scotia	0	0	Storm; Nova Scotia, 20–22 August 1950; a hurricane hit Nova Scotia, causing flooding throughout the province; many secondary roads damaged; flooding in Yarmouth, Wolfville, Windsor, Halifax, New Glasgow, and Amherst; bridges washed out. (Kindervater, 1988; Emergency Preparedness Canada database, December 1997)
1952/04	Eastend, Saskatchewan	unknown	0	Flood; Eastend, Saskatchewan; in spring 1952, the Frenchman River flooded the town of Eastend; the flood was a result of a well-above-normal snowpack in the Cypress Hills and below-normal spring temperatures. (Emergency Preparedness Canada database, December 1997)
1954/02/15	southern Ontario	unknown	0	Flood; southern Ontario, 15–18 February 1954; communities on the shores of lakes Ontario, Erie, and Huron, including Toronto, were flooded; the Thames and Grand Rivers also flooded, which disrupted utilities and forced roads and businesses to close. (Emergency Preparedness Canada database, December 1997)
1954/09/11	New Brunswick	0.052	0	Hurricane Edna; 11–13 September 1954; heavy rainfall caused severe damage and flooding in New Brunswick; highways and railways washed out; bridges damaged and destroyed; houses flooded; damage estimated at $52 000. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1954/10/15	southern Ontario	100	81	Flood; 15–17 October 1954; Hurricane Hazel hit southern Ontario, causing flooding in many communities; direct damage was estimated at $20–100 million for all of southern Ontario; 81 dead; 1868 families left homeless; flooding was widespread, from Lake Simcoe to Toronto and from the Niagara region to Lake St. Clair. In the Toronto area, winds reached 124 km/h; as much as 111 mm of rain fell in 12 hours in some areas; over 210 mm of rain fell over 2 days, causing severe flooding in the floodplain areas of the Don and Humber rivers and Etobicoke and Mimico creeks; the area hit hardest by the storm (in terms of rainfall) was Brampton and the worst flooding occurred along the Humber River; most severe flooding in the Toronto area in 200 years; local flood damage was estimated at $25 million; 20 bridges were destroyed or damaged beyond repair; the Humber River in Weston rose 6 m, sweeping away a full block of houses on Raymore Drive and killing 32 residents; a trailer park in Woodbridge was flooded and destroyed, killing 20 people. (Emergency Preparedness Canada database, December 1997)
1956/01/03	Nova Scotia	5.5	2	Flood; Nova Scotia; a prolonged warm spell from December 1955 to February 1956 caused province-wide flooding from snowmelt and ice jams; the major flooding occurred from January 3 to 18, during a period referred to as 'the longest January thaw to date'; 100 bridges were destroyed across the province; 350 houses flooded along rivers and lakes; two deaths, one by drowning, the other road related; $3.5 million in property damage, $2 million in damage to highways and bridges. (Kindervater, 1988; Emergency Preparedness Canada database, December 1997)
1956/08/30	southern Ontario	0.25	0	Flood; southern Ontario, 30–31 August 1956; flooding caused several roads to close and damage along the Thames and Grand rivers and in the Toronto area. (Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1957/08	Thetford Mines, Quebec	2	0	Flood; Thetford Mines, Quebec; a convective storm brought torrential rainfall to the area near Thetford Mines; severe damage was caused in Thetford Mines and the Bécancour River basin, totalling about $2 million. (Emergency Preparedness Canada database, December 1997)
1960/04/18	Red River, Manitoba	unknown	0	Flood; Winnipeg, Manitoba; severe flooding in the Winnipeg area of the Red River valley. (Emergency Preparedness Canada database, December 1997)
1960/05/09	southern and central Ontario	1.5	0	Flood, southern and central Ontario, 9–14 May 1960; substantial flooding caused the closing of numerous roads in central Ontario; direct damage estimated at $1.5 million. (Emergency Preparedness Canada database, December 1997)
1961/05/25	Saint John River basin, New Brunswick	3.5	1	Flood; Saint John River basins and northeastern and southwestern river basins, New Brunswick, 25–31 May 1961; $3.5 million (1961 dollars) in damage; heavy rainfall from a severe extratropical storm following snowmelt runoff resulted in widespread flooding across New Brunswick in late May 1961; significant damage was reported in the Saint John, Nashwaak, Miramichi, and Tobique river basins; one drowning victim. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1961/05/30	south-central Saskatchewan	unknown	0	Thunderstorm; south-central Saskatchewan; a Canadian record was set when 254 mm of rain fell within 1 hour at Buffalo Gap, Saskatchewan, on 30 May 1961; the rain caused road washouts, field erosion, and loss of all vegetation from several crop and pasture fields; in some areas of the storm, heavy hail fell and remained in drifts for several days. (Emergency Preparedness Canada database, December 1997)
1961/08/31	Timmins, Ontario	unknown	5	Flood; Timmins, Ontario; a severe thunderstorm moved across the Timmins area; hail was reported; storm runoff from the Town Creek basin and runoff from streets and storm sewers of Timmins caused Town Creek to reach flood stage; Timmins was flooded, resulting in road washouts, smashed houses, and undercut foundations; a mother and four children drowned in a house along the bank of the creek. (Emergency Preparedness Canada database, December 1997)
1961/10/10	northwestern British Columbia	unknown	0	Flood; northwestern British Columbia, 10–14 October 1961; flash floods caused by storm (heavy rains, warm temperatures); Nass River; houses flooded, damaged, and destroyed at Aiyanish and Greenville; much of the winter supply of fish and meat lost at Greenville, British Columbia Red Cross Society provided aid; many bridges in area damaged or destroyed. (Septer and Schwab, 1995)
1962/03/31	southeastern New Brunswick	0.26	0	Flood; New Brunswick, 31 March–4 April 1962; caused by heavy rainfall; flooding occurred along many rivers in the southeastern portion of the province; bridges, highways, and railway lines washed out; several communities and portions thereof isolated; damage in St. John, Moncton, Sackville, Kings County; cost of damage estimated to be $260 000. (Kindervater, 1985)
1962/04	Prince Edward Island/Nova Scotia	unknown	>0.3	Flood; Prince Edward Island and Nova Scotia, early April; a frontal storm hit the south coast of Prince Edward Island, depositing 180 mm of rain on top of heavy snow; major flooding occurred, causing several washouts and damaging bridges and paved highways; total damage estimated at about $300 000. In Nova Scotia; 2–10 April 1962; flooding after a snowmelt was caused by heavy rainfall accompanying the annual spring runoff (same storm as hit Prince Edward Island); the problem was increased by ice jams; the flood affected almost the entire province. (Emergency Preparedness Canada database, December 1997; Kindervater, 1988)
1963/05/03	Hay River, Northwest Territories	unknown	0	Flood; Hay River Township, south shore of Great Slave Lake, Northwest Territories, major flooding caused by ice jamming in Old Town and Indian Village required evacuation of the community of Hay River. (Emergency Preparedness Canada database, December 1997)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1964/04/16	St. John's, Newfoundland	unknown	1	Flood; Newfoundland, 16–17 April 1964; heavy rains caused the rise of a stream at Goulds, near St. John's, Newfoundland; a two-year-old child was swept into the stream and drowned. (Kindervater, 1980; Emergency Preparedness Canada database, December 1997)
1964/09/14	Nova Scotia	unknown	0	Hurricane Ethel; Nova Scotia, 14–15 September 1964; Hurricane Ethel caused widespread flooding in Nova Scotia. (Kindervater, 1988; Emergency Preparedness Canada database, December 1997)
1965	St. Lawrence River, Quebec	unknown	20	Flood; St. Lawrence River, Quebec; flood caused by an ice jam 20 km downstream from Montréal; damage was reported by several communities and 20 people died. (Emergency Preparedness Canada database, December 1997)
1966/04/14	Red River, Manitoba	12.2	0	Flood; Winnipeg, Manitoba; severe flooding in the Winnipeg area of the Red River valley, $12.2 million in flood-disaster assistance. (Emergency Preparedness Canada database, December 1997)
1966/11/03	Baie-Comeau area, Quebec	3	5	Flood; an exceptional fall flood on the north bank of the St. Lawrence River, Quebec; the region of Baie-Saint-Paul–Baie-Comeau reported the most severe damage, although the effects of the flood reached the region of Batiscan à Sept-Îles; three people died at Pointe-aux-Anglais on the Côte-Nord, two others died at Charlevoix; damage estimated at about $3 million. (Emergency Preparedness Canada database, December 1997)
1967/10/11	Nova Scotia	unknown	0	Flood; Nova Scotia, 11–12 October 1967; heavy rainfall from a cyclonic storm caused flooding throughout the Annapolis Valley and in Dartmouth. (Kindervater, 1988; Emergency Preparedness Canada database, December 1997)
1968/02/02	southwestern and southern Ontario	0.9	0	Flood; southwestern and southern Ontario, February 1968; some flooding in areas around Kettle Creek, the Thames and Ausable rivers, the St. Clair and Niagara regions, and Bayfield and Toronto; damage estimated at $0.9 million in these areas. (Emergency Preparedness Canada database, December 1997)
1969/04	Qu'appelle River basin, Saskatchewan	1.2	0	Flood; Qu'appelle River basin, Saskatchewan, April 1969; flooding caused by rapid spring snowmelt following high fall and winter precipitation; 12 000 to 14 000 acres of agricultural land flooded; urban flooding in Moose Jaw; highways, road, and bridges damaged and train derailed by washout; damage estimated to be $1.2 million (Saskatchewan Water Resources Commission, 1972)
1969/05/02	Red River, Manitoba	unknown	0	Flood; Winnipeg, Manitoba; severe flooding in the Winnipeg area of the Red River valley. (Emergency Preparedness Canada database, December 1997)
1969/05/11	Iroquois, New Brunswick	unknown	1	Flood; spring runoff created flood conditions throughout the upper portion of the Saint John River basin, New Brunswick; a nine-year-old child drowned in a river at Iroquois, Madawaska County. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1969/05/19	Newfoundland	unknown	0	Flood; Newfoundland, 19–21 May 1969; widespread flooding caused by heavy rainfall and snowmelt; railway traffic disrupted, highways washed out, houses, and buildings inundated; several municipal agencies assisted in flood relief. (Kindervater, 1980)
1970/02/02	Saint John River basin, New Brunswick	4.63	3	Flood; Saint John River basin and northeastern and southwestern river basins, New Brunswick; 2–6 February 1970; snowmelt, ice jams, and heavy rainfall caused flooding in New Brunswick; a series of ice jams in 6 rivers destroyed 32 bridges and damaged 124 others; damage estimated at $4.63 million (1970 dollars); 3 dead — 1 person died when a barn collapsed and two others drowned. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1970/04	southwestern, Quebec	2.3	0	Flood; southwestern Quebec; 1970 was marked by two exceptional floods that caused approximately $2.3 million in damage — one of the Saint-Maurice, Mattawin, l'Assomption, and Maskinongé rivers on May 5 and the other in the Chaudière River basin on April 22. (Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1970/04/30	Red River, Manitoba	unknown	0	Flood; Winnipeg, Manitoba; severe flooding in the Winnipeg area of the Red River valley. (Emergency Preparedness Canada database, December 1997)
1970/06/01	central Ontario	2.3	0	Flood; central Ontario, 1–8 June 1970; flooding in areas along the Spanish, Wanapetei, Blind, Little White, Goulais, and Mississagi rivers and in Sault Ste. Marie; several businesses and roads were closed; damage estimated at $2.3 million; most of the damage was reported in Sault Ste. Marie and on the Spanish, Wanapetei, and Blind rivers. (Emergency Preparedness Canada database, December 1997)
1971/02/13	Nova Scotia	>2.0	0	Flood; Nova Scotia, 13–15 February 1971; heavy rains and high temperatures caused snowmelt and ice jamming; widespread flood damage; houses and businesses flooded, highways and roads inundated; damage estimated by newspapers to be over $2 million. (Kindervater, 1988)
1971/04	Montréal, Quebec	0.8	0	Flood; Quebec; flooding at the mouth of the Outaouais River on des Deux Montagnes Lake and of the St. Lawrence River from the southwest corner of Montréal Island to Lake Saint-Pierre caused approximately $800 000 in damage to Montréal and the surrounding area. (Emergency Preparedness Canada database, December 1997)
1971/05	Saint-Jean, Quebec	1.2	0	Flood; Quebec; the Richelieu River overflowed, covering the surrounding area up to 45 miles away with up to 8 ft. (2.44 m) of water; 65 per cent of the damage costs were directed to the rural sector; basements in Saint-Jean were flooded, material damage was estimated at $1.2 million. (Emergency Preparedness Canada database, December 1997)
1971/08/16	Dartmouth, Nova Scotia	5.1	0	Hurricane Beth; 15–16 August 1971; 296 mm of rain fell in Nova Scotia; flood damage estimated at $3.5–5.1 million; extensive damage to crops, sections of highways and bridges were washed out; freshwater supplies in Dartmouth, Nova Scotia, were left contaminated for days because of extensive runoff into supply Lake Antigonish (Kindervater, 1988; Emergency Preparedness Canada database, December 1997)
1971/08/27	Victoriaville, Quebec	0.25	0	Hurricane Doria caused floods in the region of Bois-Francs, Quebec; the communities most affected were Thetford Mines, Victoriaville, and Warwick; damage to roads, bridges, and crops was estimated at $250 000. (Emergency Preparedness Canada database, December 1997)
1972/04	Richelieu River, Quebec	22	0	Flood; Quebec; serious flooding throughout the province, notably along the Richelieu River, which affected land from Missisquoi Bay to Saint-Jean, from mid-April to the beginning of June, causing an estimated $22 million in damage and consequent protection measures. (Emergency Preparedness Canada database, December 1997)
1972/05	central and southwestern British Columbia	10	0	Flood; central and southwestern British Columbia; freshet flooding caused significant damage along the Fraser River; an above-average snowfall and substantial mountain snowpacks, along with high temperatures in late May, resulted in severe flooding at the end of May and into June; the Fraser River peaked three times in June, flooding both the upper and lower Fraser River valley and affecting several communities (upstream communities of Kamloops and Prince George and downstream communities of Surrey and Hope); damage amounted to $10 million. (Emergency Preparedness Canada database, December 1997)
1972/05/15	Corner Brook, Newfoundland	unknown	1	Flood; Newfoundland, 15–18 May 1972; spring runoff accompanied by rainfall caused flooding affecting communities on the Port au Prince Peninsula and area; damage to Canadian National Railway lines; in Corner Brook, a three-year-old child drowned when swept away by a swollen brook behind her home. (Kindervater, 1980; Emergency Preparedness Canada database, December 1997)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1973/04/27	Saint John River basin, New Brunswick	11.9	0	Flood; Saint John River basin and northeastern and southwestern river basins, New Brunswick; flooding from heavy snowmelt runoff caused by mild temperatures and heavy rainfall caused $11.9 million (1973 dollars) in damage from 27 April to 5 May 1973, mostly in the Saint John River basin; the Sheffield–Maugerville area (downstream from Fredericton) and Fredericton were some of the communities that were flooded; the pulp and paper industry was also affected. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1974	south and central Saskatchewan	unknown	0	Flood; south and central Saskatchewan; in the winter of 1974, near-record snowfall resulted in extensive spring flooding across the province; farms and ranchland were flooded, and roads and bridges were damaged along the Carrot and Red Deer rivers; the communities of Pierceland and Green Lake on the Beaver River were affected; Moose Jaw, Regina, and Lumsden, on the Qu'Appelle River basin, received major flood damage; in Moose Jaw, the Moose Jaw River, Thunder Creek, and Spring Creek overflowed their banks, flooding 60 city blocks (including 480 houses) and leading to the evacuation of 1400 people (18–19 April); the Wascana Creek overflowed its banks in Regina. (Emergency Preparedness Canada database, December 1997)
1974/01	southern Quebec	60	0	Flood; southern Quebec; January to June; over 300 municipalities were affected by the province-wide flooding caused by an unusually wet spring combined with excessive spring runoff; flooding centred on the Gatineau, Ottawa, Richelieu, St. Lawrence, Chateauguay, Saint-Maurice, and Chaudière rivers; over 1000 houses and 600 summer cottages were flooded; approximately 7000 people were evacuated; damage was estimated at $60 million; in Maniwaki, damage was high as 3000 residents had to be evacuated; the communities of Wakefield, Chelsea, and Pointe-Gatineau, downstream from Maniwaki, were also affected. (Emergency Preparedness Canada database, December 1997)
1974/04	central and southern Manitoba	14.5	0	Flood; Manitoba, April and May 1974; major flooding in the province caused $14.5 million in damage; severe flooding occurred in central and southern Manitoba on practically all rivers and river systems; many people had to be evacuated; although the natural discharge of water was greater than the 1966 flood, Winnipeg avoided major damage because of diversion of excess water through the Red River Floodway, the Portage Diversion, and the Shellmouth dam, which were flood-control measures set up to reduce flooding in the city. (Emergency Preparedness Canada database, December 1997)
1974/04	north-central Alberta	10	0	Flood; north-central Alberta; $10 million in damage during April and May 1974. (Emergency Preparedness Canada database, December 1997)
1974/05/02	Hay River, Northwest Territories	unknown	0	Flood; Hay River Township, south shore of Great Slave Lake, Northwest Territories, flooding caused by ice jamming required the evacuation of West Channel residents. (Emergency Preparedness Canada database, December 1997)
1974/05/17	Cambridge, Ontario	10	0	Flood; Grand River, Cambridge, Ontario; 17–21 May 1974; flooding along the Grand River caused $7–10 million in damage to Cambridge and other areas along the Grand River; many roads and businesses were closed in Cambridge; some small dams (one on the Grand River and two on the Maitland River) were damaged and utility services were disrupted in some areas. (Emergency Preparedness Canada database, December 1997)
1974/07/15	Watson Lake area, Yukon Territory	unknown	0	Storm; southeastern Yukon Territory; heavy rainfall on 15–16 July 1974, along the Alaska Highway caused sections of highway and bridges to be washed away; the Watson Lake area received 100 mm of rain; at times, 25 mm of rain fell during a 6 hour period. (Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1976	Montréal, Quebec	20	0	Flood; Quebec, spring, 1976; $20 million in damage; province-wide flooding caused by an unusually wet spring combined with heavy spring runoff affected several areas of the province, most notably the Montréal area, Trois-Rivières, the Eastern Townships, and the Outaouais. (Emergency Preparedness Canada database, December 1997)
1976/03/31	Saint John River basin, New Brunswick	2	2	Flood; Saint John River basin, New Brunswick; some river systems in northeastern New Brunswick were also affected; 31 March–5 April 1976; cyclonic storms and high snowmelt contributed to cause severe ice-jam flooding; regional highways, municipal streets, railroads, and bridges were inundated; massive ponds formed in some residential areas; a state of emergency was declared; in the Perth–Andover area, 400 people were evacuated as the flood submerged about 200 houses, several businesses, and a junior high school; two people died — one from a trail derailment, the other drowned; $2 million (1976 dollars) in damage. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1976/04	Assiniboine River, Manitoba	4.5	0	Flood; Manitoba; floods caused $4.5 million in damage in April and May; record flooding on the Assiniboine River. (Emergency Preparedness Canada database, December 1997)
1976/05/04	Grand Falls, Newfoundland	0.5	2	Flood; Newfoundland; spring runoff caused the Exploits River to rise, inundating a section of highway west of Grand Falls; highway repair costs were at $500 000; two people were drowned when their vehicle went off the highway and into 3 m of water. (Kindervater, 1980; Emergency Preparedness Canada database, December 1997)
1976/08/10	New Brunswick	unknown	0	Hurricane Belle; New Brunswick, 10–11 August 1976; flooding reported across New Brunswick due to Hurricane Belle. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1976/08/27	Toronto, Ontario	1	0	Flood; Toronto, Ontario, August 1976; at least five roads were closed due to flooding; damage estimated at $1 million. (Emergency Preparedness Canada database, December 1997)
1977/03/13	southern and central Ontario Ontario	0.9	0	Flood; southern and central Ontario, March 1979; flooding in areas along the Saugeen and Maitland rivers in southwestern Ontario and in Sault Ste. Marie; direct damage estimated at over $900 000; many roads were closed along the Saugeen and Maitland rivers. (Emergency Preparedness Canada database, December 1997)
1977/12/27	Corner Brook, Newfoundland	0.1	0	Flood; across insular Newfoundland; the combination of rain and high temperatures on Christmas Day and Boxing Day melted the accumulated snow; damage to city streets in Corner Brook was estimated at $100 000. (Kindervater, 1980; Emergency Preparedness Canada database, December 1997)
1978/04	Newfoundland	5.2	0	Flood; Newfoundland; bridge on Codroy River partly destroyed; damaged roads in Corner Brook; highways flooded, roads washed out; $5.2 million in damage. (Kindervater, 1980; Emergency Preparedness Canada database, December 1997)
1978/07/10	Edmonton, Alberta	unknown	0	Flood; Edmonton, Alberta; 10–11 July 1978; a highly localized rainstorm and subsequent runoff exceeded the capacity of Edmonton's sewer system, causing a severe urban flood. (Emergency Preparedness Canada database, December 1997)
1978/09/07	Kicking Horse Pass, British Columbia	unknown	0	Flood; Kicking Horse Pass, on the British Columbia and Alberta border, 7 September 1978; debris flows triggered by a jökulhlaup from Cathedral Glacier destroyed three levels of Canadian Pacific Railway track, derailing a freight train; sections of the Trans-Canada Highway were also buried. (Emergency Preparedness Canada database, December 1997)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1978/10/29	northwestern British Columbia	>6.2	0	Flood; northwestern British Columbia, 29 October–7 November 1978; two episodes of flooding hit some parts of region, the first due to a major storm system that struck from 29 October to 1 November, the second due to heavy rainfall and a mixture of snow and rain from 4 to 7 November; flooding occurred along the Skeena and Nass rivers and on the Queen Charlotte Islands; Terrace–Kitimat region declared a disaster area; local economy severely disrupted by flooding and loss of communication links; Lakelse Lake and Old Remo areas worst hit by flooding; dozens of houses flooded; flooding also reported in Thornhill, Kitamaat Village, Terrace, and Kitimat; numerous bridges damaged and destroyed and roads/highways washed out in many places; railway line between Smithers and Terrace closed because of flooding; total cost of damages to highways $6.2 million (as reported under the November 4 to 7 flood). (Septer and Schwab, 1995)
1979/03/05	southwestern Ontario	3.5	0	Flood; southwestern Ontario; 5–8 March 1979; some rivers including the Thames, Saugeen, and Grand overflowed and caused flooding; over $3.5 million in direct damage estimated due to flooding; many roads closed along the Thames River and several businesses were closed on the Grand River; disrupted utility service in some communities on the Thames, Saugeen, and Grand rivers. (Emergency Preparedness Canada database, December 1997)
1979/03/26	Saint John River basin, New Brunswick	2.1	1	Flood; New Brunswick; snowmelt, rainfall, and ice jams caused flooding in the southern portion of the Saint John River basin, 26 March–5 April 1979; houses inundated; roads flooded and washed out; one person drowned; $2.1 million in damage. (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)
1979/04	Red River, Manitoba	18.6	0	Flood; Manitoba; April and May; $18.6 million in damage; on 10 May 1979, a major flood, close to flood levels of the 1950 flood, hit the Red River area; most of the damage occurred in the Red River valley; no communities were flooded as the dykes protecting them were temporarily raised; 10 000 people were ordered to evacuate (1450 residents were evacuated in the town of Morris and 400 from the Roseau Reserve); livestock and poultry were also evacuated. (Emergency Preparedness Canada database, December 1997)
1979/04/24	New Brunswick	2.0	0	Flood; New Brunswick, 24 April–7 May 1979; rainfall and heavy snowmelt caused extreme flood conditions in most parts of the province; several major highways closed because of flooding; houses inundated; many residents evacuated; $2 million (1979 dollars) in damage. (Kindervater, 1985)
1979/04/24	central Ontario	12	0	Flood; central Ontario, 24–30 April 1979; flooding in areas around Lake Nipissing and along the Mississagi, Blind, and Mattagami rivers; direct damage estimated at over $12 million, with $10 million in damage reported around Lake Nipissing, $1 million on the Mississagi River, and $900 000 on the Blind River; several roads were closed near Lake Nipissing and the Mattagami River. (Emergency Preparedness Canada database, December 1997)
1979/05/03	Dawson, Yukon Territory	unknown	0	Flood; Dawson, Yukon Territory; an ice jam on the Yukon River caused water to rise 2 m, flooding 80 per cent of Dawson's buildings. (Emergency Preparedness Canada database, December 1997)
1979/05/12	central Ontario	2.5	0	Flood; central Ontario, 12 May 1979; the White, Mississagi, Goulais, Spanish, and Wanapetei rivers were some of many that overflowed and caused flooding; direct damage estimated at over $2.5 million; $2 million in damage reported along the White River alone; roads and businesses were closed and utility services were disrupted in the community of White River. (Emergency Preparedness Canada database, December 1997)
1979/09/06	New Brunswick	0.5	0	Tropical storm 'David'; New Brunswick, 6–7 September 1979; worst damage in the Moncton area; total damage caused by heavy rainfall and flooding estimated at $0.5 million (1979 dollars). (Kindervater, 1985; Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1979/09/13	southern Quebec	unknown	0	Hurricane Frederic; southern Quebec; Frederic brought more than 80 mm of rain to southern Quebec. (Emergency Preparedness Canada database, December 1997)
1980/03/21	southern Ontario	11	0	Flood; southern Ontario, 21–22 March 1980; severe flooding in southwestern Ontario, especially in the Ganaraska region and the Lower Trent River; several businesses and roads were closed; an estimated $11 million in direct damage occurred along several river systems as well as in Toronto. (Emergency Preparedness Canada database, December 1997)
1980/07/17	Charlottetown, Prince Edward Island	unknown	0	Thunderstorm; Charlottetown, Prince Edward Island; a thunderstorm passed over Charlottetown; the city was flooded by heavy rains and blacked out for several hours. (Emergency Preparedness Canada database, December 1997)
1980/09/01	Saint-Hubert, Quebec	unknown	2	Flood; Saint-Hubert, Quebec; flooding due to a severe thunderstorm that dropped 40 mm of rain in 30 minutes on Saint-Hubert (east of Montréal); two deaths reported due to the flooding. (Emergency Preparedness Canada database, December 1997)
1980/12/26	southwestern British Columbia	13	0	Flood; Squamish River, British Columbia, December, 1980; a combination of heavy rain and melting snow caused flooding along the Squamish River and other nearby rivers; flooding of towns such as Brackendale resulted in $13 million in damage. (Emergency Preparedness Canada database, December 1997)
1981	Edmonton, Alberta	unknown	4	Flood; Edmonton, Alberta; a thunderstorm dropped 56 mm of rain in less than 3 hours, causing severe and widespread flooding; the flooding resulted in washouts, mud slides and four deaths. (Emergency Preparedness Canada database, December 1997)
1981/02/18	southeastern Ontario	0.9	0	Flood; southeastern Ontario; 18–20 February 1981; flood damage in the Ottawa and Rideau river basins in southeastern Ontario and in Toronto; $0.9 million in direct damage reported along the Rideau River. (Emergency Preparedness Canada database, December 1997)
1983/01/14	Newfoundland	34	0	Flood; central, southern, and western Newfoundland; 11–19 January 1983; flooding triggered by mild temperatures and excessive rainfall from a major winter storm caused damage to houses and highways along the Exploits and Gander rivers (central), the Conne River (southern), and in Stephenville (west coast); the flood occurred after 189 mm of rain fell within a 36 hour period, destroying several houses, businesses, and other infrastructure worth approximately $4 million; several houses, two railway cars, and a local club were washed away; approximately 500 people were evacuated; numerous washouts along many central highways forced travellers to seek refuge; no loss of life or personal injury was reported; flooding in the Exploits and Gander river basins damaged part of the dam and power house in Bishop Falls on the Exploits River; total damage estimated at $34 million. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1983/02/27	Badger, Newfoundland	0.089	0	Flood; Badger, Newfoundland; 27 February–10 March 1983; severe flooding in Badger due to ice and debris blockages on the Exploits River; 34 houses evacuated; town hall, firehall, and arena flooded; declared a disaster area; damage costs estimated at $89 000. (Environment Canada, 1988)
1983/03/19	southern Quebec	1.5	0	Flood; north shore of St. Lawrence River and Yamachiche and Maskinongé river basins, Quebec, 19–20 March 1983; ice jams caused flooding in Louiseville and Yamachiche, which led to road closures and the evacuation of several hundred houses, businesses, schools, and a hospital; property damage at Louiseville and Yamachiche reached $1.5 million. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1963/05/10	Montréal area, Quebec	7.9	0	Flood; Montréal area, Quebec: Ottawa River, des Prairies River, and des Mille Îles River, 10 May 1983; heavy rains caused several rivers to overflow their banks; damage approximately $4.8 million along the des Mille Îles River and $3.1 million along the des Prairies River. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1983/05/10	southwestern Quebec	2.5	0	Flood; upper Richelieu River, Quebec; several communities along the Ottawa, Gatineau, du Lièvre, and Rouge rivers were also affected; heavy rains (70–110 mm) caused widespread flooding along the upper Richelieu River on 10 May 1983; from the city of Saint-Jean to the Canada–United States border, 640 houses and 373 cottages were flooded; approximately 160 people were evacuated; the flood caused $2.5 million in damage in Canada and $3.3 million in damage in the United States. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1983/06/24	Saskatchewan/ Alberta	25	0	Storm; Regina, Saskatchewan, and Edmonton, Alberta; the storm caused flooding and $25 million (total) in damage. (Emergency Preparedness Canada database, December 1997)
1983/07/08	Regina, Saskatchewan	60	0	Storm; Regina, Saskatchewan; a storm hit Regina, 8–9 July 1983; at the airport, records show that Regina received 50 mm of rain in 1 hour and 117 mm in a 24 hour period, whereas northern Regina received up to 150 mm, with 83 mm coming in 1 hour; the storm overloaded the sewer system and caused severe flooding, mostly in basements; the floods caused an estimated $60 million in damage. (Emergency Preparedness Canada database, December 1997)
1984/01/04	Lower Fraser River valley, British Columbia	4	0	Flood; Lower Fraser River valley, British Columbia, 4–5 January 1984; heavy rainfall and snowmelt caused flooding in southwestern British Columbia; transportation was disrupted as debris jams and water obstructed traffic on Highway 1; flooding and washouts caused several problems; there was a mudslide and a bridge washout, and sections of the Canadian Pacific and Canadian National railway tracks were washed out; initial estimate of damage $4 million. (Emergency Preparedness Canada database, December 1997; Environment Canada, 1988)
1984/02/04	Newfoundland	0.1	0	Flood; western and eastern Newfoundland, 4–16 February 1984; torrential rain, warm temperatures, and ice jams caused flooding in a number of Newfoundland communities, including Stephenville, Black Duck, and Port au Port in the west and Glovertown in the east; families were evacuated and roads were closed; outgoing flights were delayed at St. John's Airport and Gander International Airport; unofficial estimate of damage $100 000. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1984/02/05	Truro, Nova Scotia	unknown	1	Flood; central and eastern Nova Scotia; 5–8 February; a combination of mild temperatures, rainfall, snowmelt, and ice jams caused flooding in the Truro area; along the Salmon and North rivers, several houses were inundated; February flood described as the worst in 50 years; one person died from fumes while pumping out a basement. (Environment Canada, 1988; Kindervater, 1988; Emergency Preparedness Canada database, December 1997)
1984/02/13	southern Ontario	0.5	2	Flood; southern Ontario; 13–16 February; flooding in Toronto, Chatham, Dresden, Port Bruce, St. Thomas, and Brantford as a result of very mild weather and rainfall; ice jams formed on several rivers, causing them to overflow; several roads were closed due to flooding, most of them along the St. Clair and Thames rivers in southwestern Ontario; in Toronto, the Don Valley Parkway was closed in areas due to overflow along the Don River; two people died as they were swept away in Black Creek in Toronto; over 100 people were evacuated in southern Ontario; utilities were disrupted in some areas and businesses were forced to close; direct damage estimated at $500 000. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1984/04/25	Saint John River basin, New Brunswick	unknown	1	Flood; Saint John River basin, New Brunswick; 25 April–6 May 1984; because of rain, mild temperatures, and snowmelt in late April and early May, the Saint John River flooded areas between Fredericton and Gagetown; several houses were inundated and the Trans-Canada Highway was closed for several days; one person was reported to have drowned when his vehicle left a flooded section of Route 690 near Lakeville Corner and plunged into deeper water. (Kindervater, 1985; Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1984/06	Winnipeg, Manitoba	unknown	0	Flood; Winnipeg, Manitoba; mid-June 1984; heavy rainfall. (Emergency Preparedness Canada database)
1984/10/06	Pemberton area, British Columbia	4.375	1	Flood; Pemberton area, southwestern British Columbia, 6–8 October 1984; persistent heavy rains over three days caused severe flooding in the Pemberton, Squamish, and Lion's Bay areas; 300 people were evacuated; the flood caused one death and an estimated $4.375 million in damage. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1985/02/23	Chatham, Ontario	0.8	0	Flood; Chatham, Ontario, and surrounding area; 23 February–1 March 1985; widespread flooding in the city of Chatham and on agricultural land in Dover and Raleigh townships downstream from Chatham as a result of mild temperatures causing snowmelt; the Thames River and McGregor and Indian creeks overflowed because of snowmelt runoff; an estimated 1180 houses were flooded or inaccessible; a dam was damaged; an Ontario Provincial Police helicopter was used to evacuate 30 people in outlying areas; $0.8 million in damage was reported on the Thames River. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1985/05/07	Hay River, Northwest Territories	0.618	0	Flood; Hay River Township, south shore of Great Slave Lake, Northwest Territories, 7 May 1985; record high flows of the Hay River and ice jams caused serious flooding, requiring the evacuation of West Channel residents; cost of damage $618 000. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1986/03/30	Chaudière River, Quebec	unknown	0	Flood; Chaudière River, Quebec; 30 March–1 April; communities along the Chaudière River were flooded during the 1986 spring thaw; heavy rain and ice jams contributed to flooding, which forced about 750 people to evacuate; an estimated 460 businesses and houses were flooded along the river; the communities of Sainte-Marie, Scott, Breakeyville, Saint-Étienne, and Saint-Lambert were directly affected by the flood. (Emergency Preparedness Canada database, December 1997; Environment Canada, 1988)
1986/04/01	Woodstock area, New Brunswick	unknown	0	Flood; Saint John River, between Simonds and Woodstock, New Brunswick; 1–4 April 1986; flooding occurred along sections of the Saint John River in early April 1986 because of ice jams; unusually warm weather contributed to rapid snowmelt and to the breakup of river ice; 37 houses in Simonds were damaged and families had to be evacuated; a private airstrip was severely damaged and three light aircraft were submerged. (Environment Canada, 1988)
1986/05	Manitoba	unknown	0	Flood; Ste. Rose du Lac, Manitoba, 5–8 May 1986; caused by snowmelt and heavy rainfall. (Emergency Preparedness Canada database, December 1997)
1986/05/16	Winisk, Ontario	7	2	Flood; Winisk, Winisk River, northern Ontario; precipitation was nearly three times normal in the western James Bay area of northern Ontario in March and April; as the snowpack melted, ice jams formed on the Winisk River; the river then overflowed its banks and Winisk was inundated; the isolation of the community, the magnitude of the flooding and a blizzard hampered rescue efforts. Almost all of the 131 residents were airlifted to safety; two died; the community was virtually destroyed; a new town site was developed upstream at Peawanuk. (Emergency Preparedness Canada database, December 1997)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1986/07/15	central Alberta	30	0	Flood; central Alberta, 15–19 July 1986; 1500 houses damaged, costs in excess of $30 million. (Emergency Preparedness Canada database, December 1997)
1986/07/16	Saskatchewan–Alberta	28.3	1	Flood; McLeod and Pembina rivers, Alberta and Saskatchewan, 16–24 July 1986; on 18 July, a flood occurred in west-central Alberta on the North Saskatchewan River, on several of its tributaries, and on tributaries of the Athabasca River; eventually Edmonton, Alberta (19 July), and Battleford and Prince Albert, Saskatchewan (22 and 24 July, respectively), were also flooded; the floods resulted from high precipitation in the mountains, foothills, and Great Plain regions, compounded by high antecedent soil moisture conditions; crops were destroyed and farm buildings damaged; hundreds of houses were flooded and severe erosion occurred within communities, causing road and bridge damage and several washouts; the extensive flooding also caused an overflow of several municipal sewer systems; one death; estimated damage $28.3 million. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1987/03/31	Perth–Andover, New Brunswick	11	0	Flood; Perth–Andover, New Brunswick, 31 March–4 April 1987; serious damage occurred along the Saint John River when an ice jam formed during snowmelt and heavy rainfall caused severe flooding; the railway bridge spanning the river was destroyed; a state of emergency was declared; mass evacuations were necessary; public and private property damage estimated at $11 million. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1987/04/01	south-central Quebec	2	0	Flood; south-central Quebec; late March–early April, 1987; heavy rains and melting snow caused spring flooding; damage was widespread because of high water levels and landslides that were brought on by the heavy rain; transportation was disrupted and hundreds of people were evacuated; most damage occurred on April 1 with conditions returning to normal 2–3 April; the Beauce region south of Québec was hit hard, as main roads were closed and houses were evacuated; damage estimates for the Beauce region were $10 million; some landslides occurred on the north shore of the St. Lawrence River east of Québec experienced some landslides; damage estimates in this area reached $2 million. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1987/07/14	Montréal, Quebec	94	2	Flood; Montréal, Quebec; 14 July 1987; severe thunderstorms and intense rainfall with 100 mm of rain in 1 hour; flooding caused an estimated $94 million in damage; major expressways were flooded with over 3 m of water; in 30 minutes, the subterranean Decarie Expressway was flooded with 3.6 m of water; the sewer system was backed up, affecting approximately 40 000 houses and businesses; the subway system was closed; two people died — one from drowning in his car and the other from electrocution; damage from flooding $94 million. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1987/07/31	Upper Peace River, British Columbia/Alberta	5	0	Flood; Upper Peace River district, Alberta and British Columbia, 31July–4 August 1987; heavy rainfall produced extensive flooding, affecting tributary streams of the Peace River in northeastern British Columbia and in Alberta between 21 and 27 July; in British Columbia, roads were washed out, campgrounds had to be evacuated; in Alberta, transportation facilities were extensively damaged; a 96-car freight train derailed when it ran through a washed out section of Canadian National Railway track; extensive areas of cropland in Alberta were damaged by the high water; initial estimates of damage $5 million. (Environment Canada, 1988; Emergency Preparedness Canada database, December 1997)
1988/06	Lesser Slave Lake, Alberta	15	2	Flood; Lesser Slave Lake, Alberta; June 1988; two dead; $15 million in damage. (Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)[a]	Deaths	Comment
1988/06/29	Mackenzie Valley, Northwest Territories	6	0	Flood; two severe flood events (29 June–3 July and 11–15 July) in the Laird River and Mackenzie River basins of southwestern Northwest Territories caused $6 million in damage, mainly to highways and the Interprovincial Pipe Line Ltd. oil pipeline between Norman Wells and Zama Lake. (Emergency Preparedness Canada database, December 1997)
1989	Leamington, Ontario	15	0	Rainstorm; Kent, Essex, Leamington counties of Ontario; a severe rainstorm caused flooding in some areas and $15 million in damage. (Emergency Preparedness Canada database, December 1997)
1989	Attawapiskat River, Ontario	unknown	0	Flood; Albany and Attawapiskat rivers flood, northern Ontario; 1000 evacuated. (Emergency Preparedness Canada database, December 1997)
1989/07/19	Harrow, Ontario	35	0	Flood; Harrow and Colchester South, Essex County, Ontario; 19–20 July 1989; widespread flooding after 450 mm of rain fell in a 30 hour period; the rainfall was from an intense thunderstorm that was centred over Colchester South and seemed to be stationary over Harrow for most of its duration; the storm and floods caused $35 million in damage to agricultural areas including washouts and erosion; many people were evacuated from areas of extreme flooding and erosion; a few injuries were reported, no deaths, utility service in some areas was disrupted. (Emergency Preparedness Canada database, December 1997)
1990	Newfoundland	2.9	0	Flood; Newfoundland; floods caused $2.9 million in damage; federal financial assistance was required. (Emergency Preparedness Canada database, December 1997)
1990/05	High River, Alberta	unknown	0	Floods; heavy rains affected a number of communities in Alberta during several separate periods between May and July 1990; flooding caused damage to public works and farmland and extensive damage to private residences; communities affected included Thorsby, Winfield, Calman, Mudhurst, and Bentley. (Emergency Preparedness Canada database, December 1997)
1990/06/13	southern British Columbia	10	4	Flood; southern British Columbia; washouts and mudslides closed sections of highways 5, 33, and 97 and reduced the Trans-Canada Highway to one lane near Revelstoke; four houses were destroyed by a mudslide 16 km east of Enderby; three houses were destroyed at Stoner Creek; two people died when caught in a mudslide 10 km south of Vavenby; two others died in a house that was destroyed by a mudslide 30 km southeast of Kelowna; damage costs were predicted to be $10 million. (Emergency Preparedness Canada database, December 1997)
1990/11/08	southwestern British Columbia	5 to 10	0	Flood; southwestern British Columbia, including Vancouver Island, 8–13 November 1990; 309 people evacuated and 8 houses destroyed; heavy rains caused widespread flooding over a wide area of southwestern British Columbia and Vancouver Island; the British Columbian interior was cut off from the coast for several days as highways were shut down by landslides and road washouts; damage was estimated to be between $5 and $10 million. (Emergency Preparedness Canada database, December 1997)
1991/04/07	Chaudière River, Quebec	5 to 10	0	Flood; residents along the Chaudière River, south of Québec, experienced the highest flood levels in several years during the weekend of 7–8 April; many houses and businesses were damaged by water and ice; preliminary estimates of damage at between $5 and $10 million; 1000 residents were evacuated. (Emergency Preparedness Canada database, December 1997)
1991/10/08	northwestern British Columbia	3.7	0	Flood; northwestern British Columbia, 8–14 October 1991, caused by heavy rainfall; flooding along the Kitimat, Skeena, and Nass rivers; flood damage in Terrace, Stewart, and Prince Rupert; highways and forest service roads flooded and washed out, numerous bridges damaged and destroyed; communities along Nass River isolated; fresh food flown in through the Provincial Emergency Program; erosion damage to spawning beds; damage costs to highways, forest service roads, and a pipeline totalled $3.7 million. (Septer and Schwab, 1995)

Appendix A (cont.)

Date (y/m/d)	Location	Damage ($M)*	Deaths	Comment
1992/02/28	Peace River, Alberta	3	0	Flood; Peace River, Alberta, 388 km northwest of Edmonton, 28 February–4 March 1992; unseasonably warm weather broke up the frozen Peace River, causing water levels to rise 4.3 m above average; extensive flooding necessitated the evacuation of approximately 3800 residents of the town for at least 3 days; approximately 20 houses sustained severe damage, 30 others were damaged; no deaths or injuries were reported. (Emergency Preparedness Canada database, December 1997)
1992/03/12	Montmorency River, Quebec	unknown	0	Flood; Sainte-Brigitte-de-Laval, Quebec, about 40 km north of Québec; approximately 40 people were airlifted from this community on île Enchanteresse when flood waters caused by an ice jam on the Montmorency River rapidly inundated the town, flooding 80 houses; two Quebec government helicopters and one Canadian Forces helicopter were used in the evacuation. (Emergency Preparedness Canada database, December 1997)
1992/04/25	Hay River, Northwest Territories	unknown	0	Flood; Hay River, Northwest Territories; approximately 100 residents were evacuated as a 10 km long ice jam at the mouth of the Hay River caused backflooding in the town. (Emergency Preparedness Canada database, December 1997)
1992/05/18	Attawapiskat, Ontario	unknown	0	Flood; Attawapiskat, northern Ontario; approximately 1000 residents of this remote Cree village near James Bay were evacuated when the Attawapiskat River flooded. (Emergency Preparedness Canada database, December 1997)
1992/07/31	Toronto, Ontario	5	0	Flood; Toronto, Ontario; flood caused an estimated $5 million in damage. (Emergency Preparedness Canada database, December 1997)
1992/08/28	Elmira, Ontario	4	0	Flood; Elmira, Ontario; floods caused an estimated $4 million in damage. (Emergency Preparedness Canada database, December 1997)
1993/07/04	Swan River, Manitoba	20	0	Flood; Manitoba; prolonged rains resulted in severe flooding in the rural municipalities of Swan River, Minitonas, and Ethelbert, the local government district of Mountain, and adjoining municipalities, causing extensive damage to houses, public utilities, and municipal infrastructure. (Emergency Preparedness Canada database, December 1997)
1993/07/25	Winnipeg, Manitoba	500	0	Flood; Manitoba, 25 July–14 August 1993; Winnipeg was declared a disaster area because of flooding caused by prolonged heavy rainfall; three severe rainstorms in a five-week period during July caused sewer backups and other extensive damage to houses, power lines, and agricultural land infrastructure in Winnipeg and surrounding areas including Selkirk, Springfield and Beausejour. (Emergency Preparedness Canada database, December 1997)

Date (y/m/d)	Location	Damage ($M)*	Deaths	Comment
1996	Flat Bay, Newfoundland	unknown	0	Flood; Flat Bay, east coast of Newfoundland; flooding caused extensive damage; federal financial assistance was requested. (Emergency Preparedness Canada database, December 1997)
1996/07	southern Quebec	1000	10	Flood; widespread flooding in southern Quebec, the worst occurring in the Lac-Saint-Jean–Chicoutimi–La Baie area, Saguenay River valley; 19–21 July 1996; up to 270 mm of rain fell in 36 hours; dams overtopped, damaged, and in some cases, breached; road and railway bridges damaged and destroyed; roads washed out; houses and businesses flooded; extensive bank erosion; damage in the region estimated to exceed $1000 million with 10 deaths due to washed out roads and landslides; most costly Canadian flood event to the date. (Emergency Preparedness Canada database, December 1997)
1996/11/29	Nechako River, British Columbia	unknown	0	Flood; Nechako River, British Columbia; record low temperatures on 21 November 1996, caused thick ice formation on the Nechako River which resulted in ice-jam flooding; an ice jam 5 km long raised water levels along the low-lying regions of Prince George, flooding industrial and residential areas; parking lots, storage yards, and lumber mill flooded; ice jams caused intermittent flooding at Vanderhoof and Prince George. (Emergency Preparedness Canada database, December 1997)
1997/01/09	Durham, Ontario	unknown	0	Flooding; Durham, Ontario; on 9 January the mayor of Durham declared a state of emergency due to flooding of the Saugeen River; approximately 200 residents evacuated; some businesses and schools closed; sewage backup; relief efforts continued for over one week because of poor weather conditions; most evacuees had returned by January 24. (Emergency Preparedness Canada database, December 1997)
1997/05	Red River, Manitoba	500	3	Flood; Winnipeg; the Assiniboine, Red, and Winnipeg rivers, May 1997; spring melting of a heavy snow pack caused widespread and prolonged flooding along the Red River valley in North Dakota, Minnesota, and southern Manitoba. In Canada, over 7000 military personnel were employed for 36 days to assist in preventing flood damage and in relocating 25 450 evacuees; communities affected included Emerson, Rosenort, St. Jean Bapiste, Morris, St. Adolphe, Nutimik, Portage la Prairie, Ste. Agathe, St. Norbert, Letellier, Dominion City, Scanterbury, Halstadt, Brunkild, Domain, Sanford, and Winnipeg; three people drowned; cost $815 million. (Emergency Preparedness Canada database, December 1997)
1997/05	central British Columbia	14	0	Flooding; Terrace, Kamloops, Prince George, and Nelson areas; localized flooding in several communities; spring flood damage costs estimated at $4 million, response costs at $10 million. (Emergency Preparedness Canada database, December 1997)

Volcanoes and volcanic hazards

C.J. Hickson[1] and B.R. Edwards[2]

Hickson, C.J. and Edwards, B.R., 2001: Volcanoes and volcanic hazards; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 145–181.*

Abstract: In Canada, potentially active volcanoes are found only in British Columbia and the Yukon Territory, where they represent cinder cones, shield volcanoes, and stratovolcanoes. Magma composition controls how a volcano erupts. Low-viscosity magmas (basalt) erupt effusively and represent a low-level hazard. Magmas with higher viscosity (dacite) often erupt explosively, creating significant, far-reaching hazards. The explosion creates small rock fragments or ash that is often carried high above the volcano. These abrasive particles, caught in high-altitude winds, are carried for hundreds of kilometres and affect equipment with moving parts. Hot avalanches of semi-molten rock, or pyroclastic flows, can destroy large areas around the volcano. Additionally, when they descend snow-covered flanks, they can lead to catastrophic melting, sending mud and water cascading into surrounding drainage systems. Even long after an eruption, denuded slopes and unstable rock faces can make the area prone to floods and rock avalanches.

Résumé : Au Canada, les volcans potentiellement actifs se trouvent en Colombie-Britannique et dans le Territoire du Yukon et vont des cônes de scories aux volcans-boucliers et aux stratovolcans. La composition du magma contrôle la nature de l'éruption volcanique. Les magmas très fluides (basaltiques) se déversent d'une manière effusive et représentent un niveau de risque faible. Les magmas très visqueux (dacitiques) provoquent souvent des éruptions explosives qui créent des dégâts considérables, sur une grande distance. L'explosion forme de petits fragments de roches ou de la cendre qui sont souvent projetés très haut au-dessus du volcan. Ces particules abrasives, captées par les vents de haute altitude, sont transportées sur des centaines de kilomètres et peuvent endommager le matériel muni de pièces mobiles. Les déferlements de roches semi-fondues ou les coulées de débris pyroclastiques peuvent détruire de vastes régions situées tout autour du volcan. En outre, lorsqu'ils dévalent sur les flancs enneigés, ils peuvent provoquer des fontes de neige catastrophiques, entraînant des traînées de boue et des cascades d'eau dans les réseaux hydrographiques environnants. Très longtemps après que ne se soit produit une éruption, les pentes dénudées et les faces des roches instables peuvent rendre la région inondable et provoquer des avalanches de pierres.

---

[1] Geological Survey of Canada, 605 Robson Street, Suite 101, Vancouver, British Columbia V6B 5J3
[2] Department of Geology, Grand Valley State University, 1 Campus Drive, Allendale, Michigan 49401-9403

INTRODUCTION

The world began as a fiery ball of superheated space debris 4.5 billion years ago. As the ball cooled and solidified over hundreds of millions of years, massive volcanic eruptions, fed by upwelling hot material, created terra firma, forming the continents and the surrounding oceans (Fig. 1). The slow cooling of the Earth led to a gradual diminishing of the level of volcanism, but present-day volcanoes are proof that deep below our feet the Earth is still hot and churning.

Volcanoes are not unique to the our planetary home. Many other planets and satellite bodies in our solar system have either active or extinct volcanoes. For example, much of the Moon's surface is covered by ancient lava flows. These form the dark seas or 'mare' that are visible with the naked eye. Some of the ancient volcanoes on Mars are thought to have formed under a glacial ice cap or frozen sea, much like volcanoes now found in Iceland, Antarctica, and Canada. Venus appears to be one of the most volcanically active planets and in fact may still have active volcanoes. Currently, the most volcanically active body in our solar system is Io, one of the moons of Jupiter. Its large volcanoes erupt unique, sulphur-rich magmas.

On Earth, volcanoes are a vital part of life. They continue to create new land, and volcanic eruptions release water and carbon dioxide into the Earth's atmosphere in great quantities, helping replenish these crucial gases in our atmosphere. Volcanic ash is one of nature's best fertilizers, and many of the world's most productive growing regions are on the flanks of dormant volcanoes. Other important byproducts, such as hot water for heating and bathing and steam for producing electricity, are found in volcanic regions. Light-weight volcanic rocks are used for road beds and as special aggregate in cement, creating strong, yet light-weight building material. On movie sets, many of the huge boulders lifted by Hercules are really blocks of light-weight volcanic rocks. Volcanic mud is used in drilling oil and gas wells.

Volcanoes bring to the surface of the Earth and concentrate many important and valuable resources, such as diamonds, copper, and gold. Ancient deposits found in association with volcanoes are some of the most important sources of gold throughout Canada. Ancient seafloor volcanoes contributed to massive accumulations of base metals such as lead, zinc, and copper. Seafloor volcanoes are also important in controlling the chemistry of ocean waters.

Despite these positive impacts, volcanic eruptions can directly and indirectly wreak havoc on people living nearby and travelling in aircraft, and can even temporarily change the climate worldwide. Because volcanoes can so drastically change our day-to-day living, they have been studied for hundreds of years and are the focus of intense geological and volcanological research. Volcanoes spew forth not-so-welcome gases such as sulphur dioxide that can lead to acid rain and destruction of the ozone layer. Some major volcanic eruptions can inject so much ash into the upper atmosphere that it can lead to temporary global cooling. Volcanoes are also sources of potentially hazardous trace elements that can be detrimental to health. In Canada, young volcanoes occur in British Columbia and the Yukon Territory and are part of the system of volcanoes found around the margins of the Pacific Ocean that has long been referred to as the 'Pacific Ring of Fire' (Fig. 2).

The seriousness of the hazard represented by a volcanic eruption depends on many things, but especially on the style of eruption. Volcanoes, dominated by passive, lava-forming eruptions (like those typically seen in Hawaii), generally only threaten immobile objects such as buildings, but are not a serious threat to human life. However, poisonous gases such as sulphur dioxide and fluorine are released during some

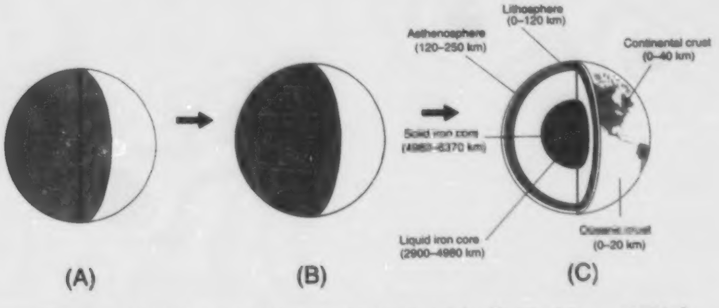


***Figure 1.** Model for the physical evolution of the Earth. After an initial stage of condensation and accretion (**A**), the planet was heated to the point where iron (Fe) was segregated into its core (**B**). Plate tectonic processes helped to form chemically distinct layers within the Earth including the crust (continental and oceanic crust have different thicknesses), the mantle, and the core (**C**). The lithosphere is the rigid outer layer of the Earth that forms the tectonic plates that move across the surface of the planet. The asthenosphere is an area in the upper mantle with small amounts of melted material that acts as a lubricant for the tectonic plates to move over.*

passive eruptions and can be life-threatening. On the other hand, large (and fortunately rare) eruptions such as those that formed the calderas underlying Yellowstone National Park in Wyoming (600 000 years ago) can have a devastating impact on literally hundreds of thousands of square kilometres of land downwind from the volcano. Such large-scale eruptions may also cause worldwide shifts in weather patterns, including a lowering of global average temperatures; it has been suggested that a huge eruption in Indonesia 74 000 years ago started the last ice age.

Canada has examples of almost every type of volcano found in the world, including stratovolcanoes, shield volcanoes, calderas, cinder cones, and maars. Although none of Canada's volcanoes are currently erupting, several volcanoes and volcanic regions are considered to be potentially active. In addition, volcanic eruptions in Alaska or along the west coast of the United States (Washington, Oregon, and California) can have a significant impact on agriculture and air travel across much of Western Canada. Large-scale eruptions anywhere in the world have the potential to affect weather patterns in Canada.

TECTONIC ENVIRONMENTS

As the Earth cooled from a gaseous ball, its interior formed several layers (Fig. 1). The innermost layers, or core and lower mantle, do not directly generate volcanic eruptions. It is the outermost layers, or lithosphere and asthenosphere, from which volcanoes form. According to the theory of plate tectonics, the Earth's crust is divided into large blocks or plates that slide around on the planet's outermost layer (Fig. 2). The theory has been almost universally accepted largely because it explains many geological events and patterns in a simple and unified way. The study of volcanoes also benefited significantly from this theory. Maps showing the boundaries of tectonic plates and volcanoes show that most volcanoes occur near the edges of tectonic plates (Fig. 2). Even more significantly, specific types of volcanoes are most commonly found along the same type of plate edge (referred to as 'tectonic environment'). When one plate moves, by necessity it must push against the neighbouring plate, pull away from the neighbouring plate, or move parallel to the neighbouring plate. These three types of relative motion form three distinct types of plate edges or boundaries. When two plates collide, one is usually forced under the other to form a subduction zone (Fig. 3). When two plates pull apart, hot material from beneath the plates rises to fill the intervening gap, forming a spreading ridge in the oceans or a rift valley on the continents. The boundary of two plates that move past each other in a parallel way is referred to as a 'transform boundary'. Volcanoes form in all three of these tectonic environments, especially the first two (subduction zones and spreading ridges).

In a subduction zone, one piece of crust is forced down into the mantle below the overriding crust (Fig. 3). In most subduction zones, the crust being forced into the mantle is oceanic crust, which is generally denser than continental crust and commonly covered in thick, wet sediments and ooze. As these sediments and ooze are carried into the Earth's mantle with the oceanic crust, rising heat and pressure dry them out in a process called 'dehydration'. The water expelled from the descending crust rises and changes the chemical composition of the overlying mantle. The presence of water actually lowers the melting temperature of the

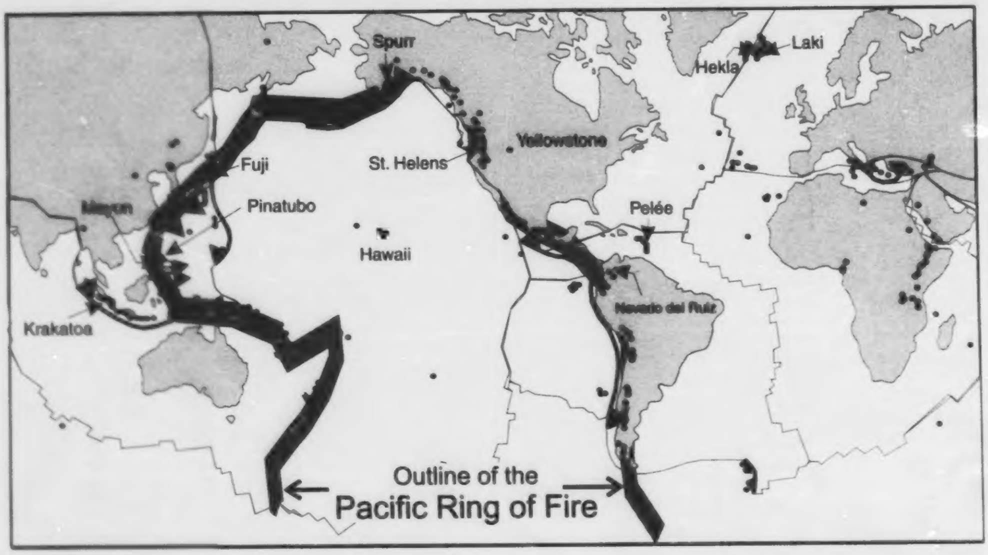


Figure 2. Distribution of many active and recently active volcanoes and boundaries of major tectonic plates. Bold lines are convergent boundaries thought to have active subduction zones. The shaded area is known as the 'Pacific Rim of Fire'. Shown are the names and locations of several volcanoes mentioned in the text (modified from *Krafft, 1993).*

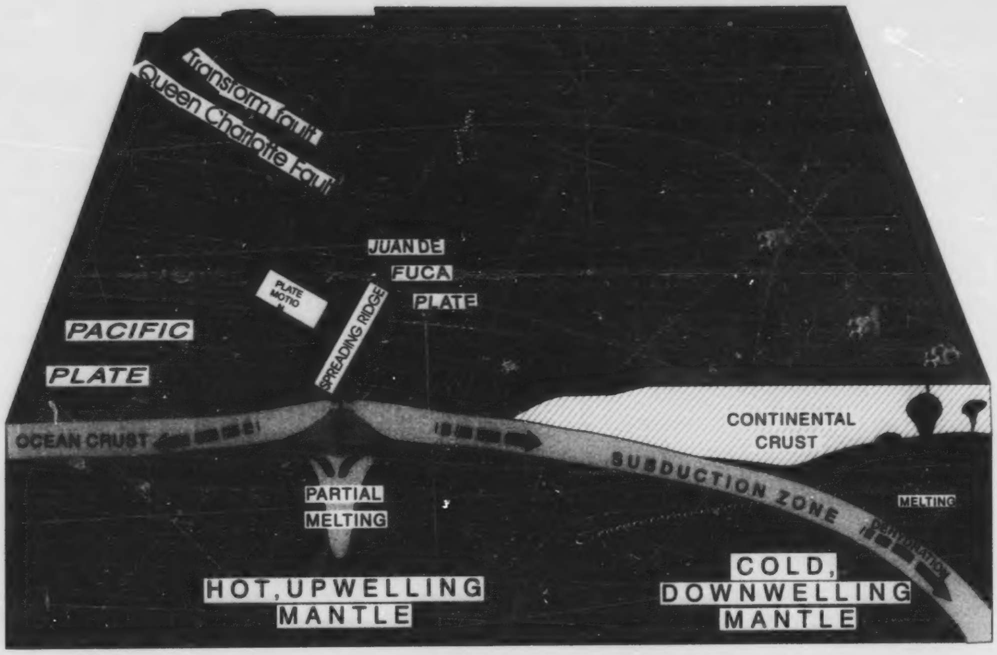


Figure 3. *Schematic block diagram showing major features of a subduction zone and spreading ridge, such as are found in southwestern British Columbia. The North American Plate is travelling westward and colliding with the smaller Juan de Fuca Plate. The collision creates earthquakes and volcanoes. The chain of volcanoes that forms is called an 'arc'. Mounts St. Helens, Rainier, Baker, Garibaldi, and Meager belong to the Cascade magmatic arc.*

overlying mantle rocks and causes them to melt. Melting mantle forms magma that rises through the crust to erupt on the surface, forming a volcano. In Canada, subduction-zone volcanoes are found only in extreme southwestern British Columbia, as part of the Cascade volcanic arc (e.g. mounts Garibaldi and Meager; *see* Fig. 3, 4, and Appendix A).

In other parts of the Earth, instead of colliding, plates move away from one another. Commonly, this occurs in the oceans, creating an oceanic ridge. Along the ridge, rising magma creates new oceanic crust. These ridges are formed mostly of basaltic lava. The Juan de Fuca Ridge off the west coast of Vancouver Island is just such a spreading ridge (Fig. 3). On continents, the crust can also be stretched and torn apart, forming regional weaknesses and , in places, large valleys. One such valley is the East African rift zone; in Western Canada, the northwestern corner of the British Columbia is undergoing such stretching and weakening of the crust. In fact, some of Canada's youngest volcanoes formed in the weakened zone. They are mostly basalt and are commonly small cinder cones (e.g. Lava Fork or Tseax cone; *see* Fig. 4 and Appendix A), though much larger volcanoes also exist, such as Mount Edziza (*see* Fig. 4 and Appendix A).

Yet another type of volcano-forming tectonic environment is found where hot, upwelling mantle forms magmas beneath an expanse of crust. The upwelling mantle material, focused on a single spot, creates a volcano. The crust continues to move, however, and a series or chain of volcanoes forms. The oldest volcanoes are found in the direction towards which the crust is moving and youngest, in the direction from which it is coming. These chains of volcanoes, called 'hot-spot volcanoes', are often large, basaltic, shield volcanoes. A famous example is the Hawaiian Islands. The oldest volcanoes have sunk beneath the waves west of Kauai, and Lohi, the newest volcano in the Hawaiian chain, is slowly rising from the ocean floor off the east coast of the island of Hawaii, although it has yet to break the surface of the ocean. Nazko cone (*see* Fig. 4 and Appendix A), in central British Columbia, forms the eastern end of a hot-spot chain of volcanoes that can be traced back 14 million years to now long-eroded volcanoes along the coast of British Columbia.

THE BUILDING BLOCKS OF VOLCANOES

All volcanoes start out the same way, as a liquid that forms beneath the surface of the Earth. These subsurface liquids are made up mostly of the following eight elements: oxygen (O), silicon (Si), aluminum (Al), iron (Fe), magnesium (Mg), calcium (Ca), sodium (Na), and potassium (K). Compounds referred to as 'volatiles' (or gases), such as sulphur dioxide, carbon dioxide, and water, are also commonly present in these liquids in minor amounts. The subsurface liquids, or silicate melts (because they contain abundant silica), are referred to as 'magmas' until they reach the surface, when they become 'lavas'. Lava flows are most commonly streams of very hot silicate liquids that can carry crystals formed underground or grow as the lava cools. Lavas vary from 500°C to more than 1200°C and can erupt in many different ways depending on variations in their chemical composition, especially in the relative amounts of silicon and water.

Volcanoes are built of rocks that record the history of eruptions at a particular volcano. Because their chemical composition varies tremendously, volcanic rocks have many different names. These names have been chosen to convey as much information as possible about the rock in a single word. Because volcanic rocks are composed mostly of the elements Si and O, for convenience in describing them, volcanologists

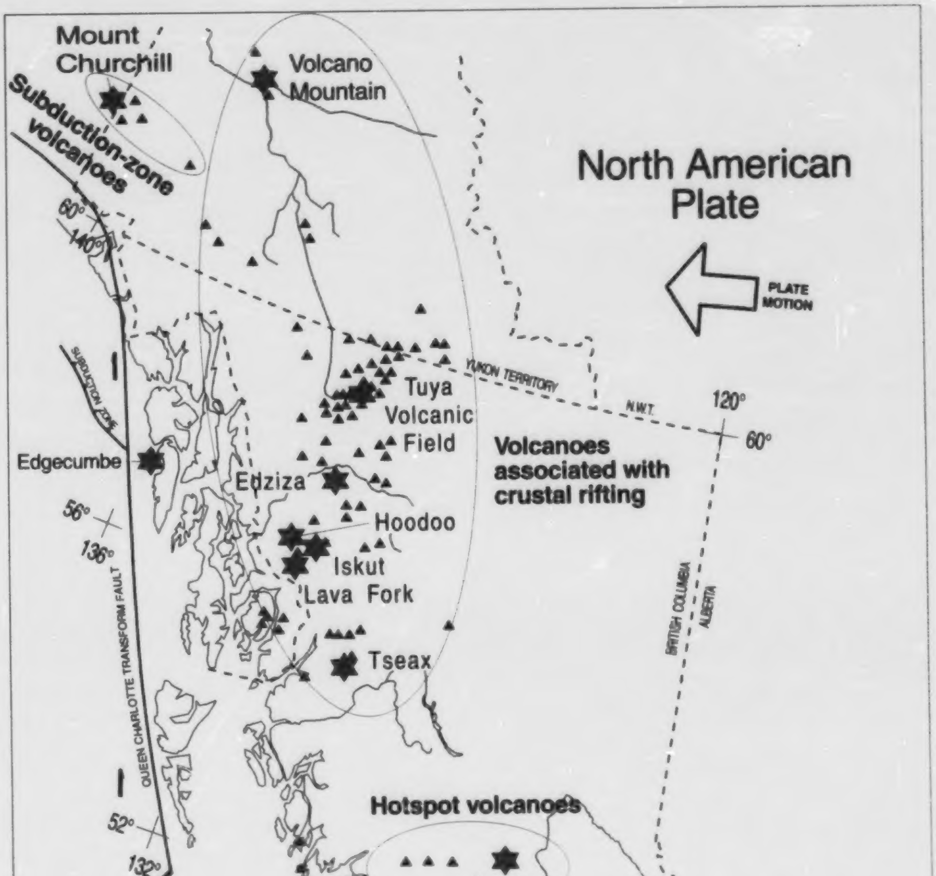

Mount Churchill
Volcano Mountain
Subduction-zone volcanoes
North American Plate
PLATE MOTION
60°
140°
SUBDUCTION ZONE
YUKON TERRITORY
N.W.T.
120°
60°
Tuya Volcanic Field
Volcanoes associated with crustal rifting
Edgecumbe
Edziza
Hoodoo
56°
136°
Iskut
Lava Fork
QUEEN CHARLOTTE TRANSFORM FAULT
BRITISH COLUMBIA
ALBERTA
Tseax
Hotspot volcanoes
52°
132°
Nazko
Wells Gray
200 km
Pacific Plate
Meager
'Explorer Plate'
Garibaldi
Subduction-zone volcanoes
PLATE MOTION
SUBDUCTION ZONE
Juan de Fuca Plate
Baker
USA

Figure 4. Quaternary (triangles) and Holocene (stars) volcanoes in Western Canada and their tectonic settings. Holocene volcanoes are discussed in detail in Appendix A.

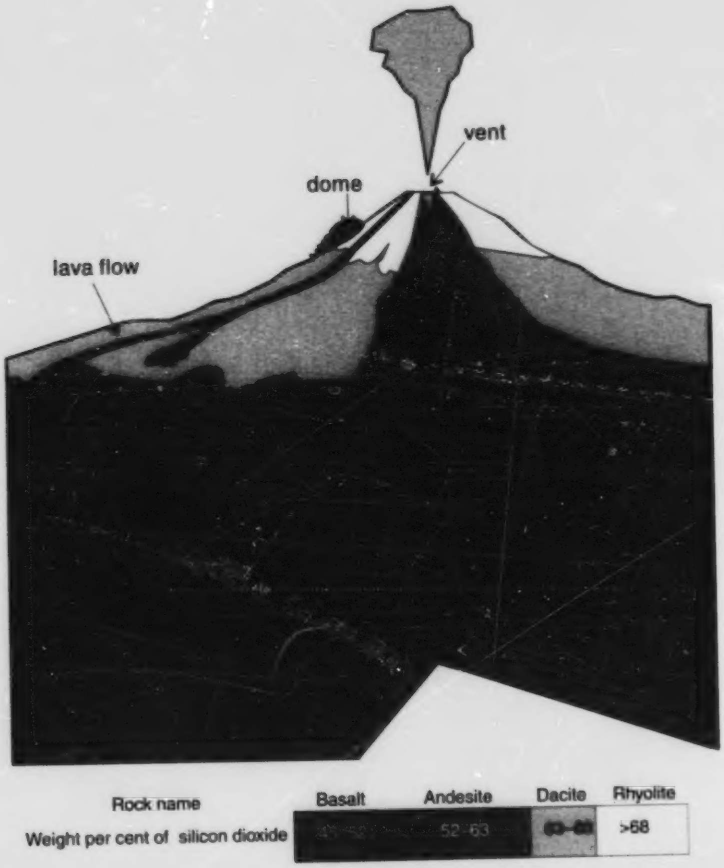


Figure 5. Schematic figure of a stratavolcano illustrating some volcanic terms and rock names. Rock names are based on the weight per cent of silicon dioxide (SiO_2) in the rock (courtesy of the Cascade Volcano Observatory, United States Geological Survey).

combine these two elements and refer to them as 'silica' (SiO_2). These two elements account for over 45% (by weight) of the rocks; thus, the names of volcanic rocks vary depending on the amount of SiO_2 present in the rock. For example, basalt, the most common volcanic rock in Canada (and on the surface of the Earth), contains between 45% and 52% SiO_2 by weight (wt. %; Fig. 5). Large volumes of basalt are found throughout Canada. In Western Canada, basalt covers much of the Chilcotin and Cariboo plateaus of central British Columbia. Andesite, another kind of volcanic rock that contains more SiO_2 than basalt (52–63 wt. % SiO_2; Fig. 5), is not as common at Canada's youngest volcanoes, but could potentially erupt in the Garibaldi area of southern British Columbia or from Mount Baker, immediately south of the border in northwestern Washington State. Basalt and andesite usually flow easily and most commonly form lava flows. Volcanoes that erupt dacite (63–68 wt. % SiO_2; Fig. 5) and rhyolite (>68 wt. % SiO_2; Fig. 5) lavas are even less common, but are potentially more explosive and dangerous than volcanoes that erupt basalt and andesite lavas. Because dacite and rhyolite magmas are thick and viscous (i.e. 'sticky', like honey or molasses), they do not flow easily and are prone to explode when they are rich in volatiles. Dacitic and rhyolitic lavas that contain little gas, build up around the vent area as rounded, muffin-shaped mounds of lava called 'domes' (Fig. 119). Two large eruptions of dacite magma, at Mount Meager in southern British Columbia and at Mount Churchill in eastern Alaska (a few kilometres west of the Yukon Territory border), produced significant deposits of volcanic ash across much of Western Canada in the last 3000 years (Fig. 6, in pocket).

Whether the magma is basaltic, andesitic, dacitic, rhyolitic, or a myriad of other, more obscure compositions, different names are used to describe it once it has erupted. Depending on its chemical composition (and other factors such as the volatile content), magma will either flow out of the volcano's vent (Fig. 5) passively and be 'coherent' (e.g. lava), or it will spew from the vent violently and be fragmental (pyroclastic). As they cool, coherent lava flows often develop very regular fractures called 'columnar joints' (Fig. 7) that are easily recognized and very distinctive. Lava flows may also produce some fragmental material, referred to as 'flow breccia', that forms as the lava cools, but continues to move. The surface of the lava can also have different forms. Ropy, shell-like surfaces are called 'pahoehoe' and broken, vesicular surfaces, 'aa' (Fig. 8). If, however, the magma breaks into fragments during an explosive eruption, the fragments are called 'tephra' or 'scoria' (Fig. 9). These fragments are described on the basis of their size and the volume percentage of vesicles (i.e. air bubbles; Fig. 10a, b) they contain. Magma fragments whose largest dimension is over 64 mm are referred to either as 'blocks' or 'bombs', depending on their shape (Fig. 9, 11). Blocks are angular and composed of magma that solidified before being thrown from the vent. Bombs, on the other hand, are ejected while still partly molten and can show signs of aerodynamic shaping during flight (Fig. 11). Magma fragments with a maximum dimension under 64 mm and over 2 mm are referred to as 'lapilli' (Fig. 9). 'Ash', the term used for fragments that are less than 2 mm in maximum dimension, is the smallest type of tephra. 'Cinder' and 'scoria' are terms used to describe tephra that is moderately vesicular (<75% vesicles by volume) and most commonly basaltic to andesitic in composition. The term 'pumice' is reserved for tephra that is highly vesicular (>75% vesicles by volume; Fig. 10b). Pumice most often forms from dacitic or rhyolitic magmas and is commonly so light that it will float on water; large pieces are easily carried.

Figure 7. Columnar joints in a solidified lava flow. These regular, polygonal joints form as a result of contraction as the lava cools. Photograph by C.J. Hickson. GSC 2000-121

Figure 8. Pahoehoe (foreground) and aa (background). Photograph by C.J. Hickson. GSC 2000-122

Figure 9. Scoriaceous tephra from the Nazko cone, central British Columbia. The smaller pieces (less than 6.4 cm long) are lapilli sized and the larger pieces are block or bombsized. For comparison, the geological hammer is approximately 30 cm long. Photograph by C.J. Hickson. GSC 2000-123

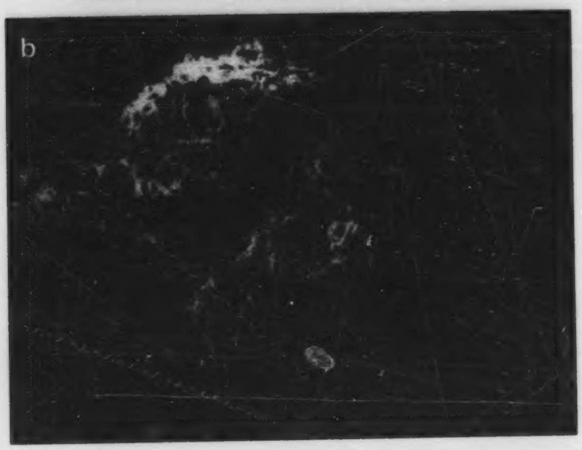


***Figure 10.** a) Vesicles in solidified lava. Photograph by C.J. Hickson. GSC 2000-124 b) Scanning electron microscope image of a pumice fragment showing its very porous, vesicular nature. Photograph by C.J. Hickson. GSC 2000-125*

***Figure 11.** Volcanoes commonly shoot out large blobs of hot magma called 'volcanic bombs'. Because the magma is very hot, as it sails through the air, it can develop a streamlined shape like that of the 'spindle bomb' shown here. Photograph by C.J. Hickson. GSC 2000-126*

TYPES OF VOLCANOES

Volcanoes and volcanic vents can have a variety of shapes. A volcanic vent is the hole from which the magma emerges from beneath the Earth's surface. Vents mark the birthplace of the volcano and are most commonly roughly circular. Some, however, are long cracks in the ground and are called 'fissures'. Volcanoes are given specific names depending on their shape (or morphology). Stratovolcanoes and shield volcanoes are the largest types of volcanoes, but have distinctly different shapes because of differences in the chemistry of the erupting magma. Shield volcanoes and stratovolcanoes erupt many times over thousands to hundreds of thousands of years whereas cinder cones, lapilli cones, and maars, which are smaller volcanoes with different morphologies, usually have a short life, erupting only once.

Classically shaped volcanoes, such as Mount Fuji in Japan with its sharp, pointed top and sloping, concave sides, are what comes to mind when we think of stratovolcanoes. Stratovolcanoes form from repeated eruptions of viscous, slow-moving lavas (most commonly andesite and dacite) and are common in subduction zones. Because the lavas are viscous, they do not move quickly and commonly flow only a few kilometres from the vent. Explosive eruptions are often associated with these volcanoes. No examples of stratovolcanoes resembling Mount Fuji are found in Canada. Many of Canada's volcanoes may have started out as Mount Fuji-like stratovolcanoes, but erosion by glacial ice and water has destroyed most of their original cone shape, leaving them as jagged, irregular peaks. Many of the more active volcanoes in Alaska and many of the Cascade volcanoes south of the Canadian–American border (e.g. Mount Baker, Fig. 12) are 'classic' stratovolcanoes.

Shield volcanoes commonly are not as dramatic-looking as stratovolcanoes, but are often much larger. The Hawaiian Islands are a series of large, gently sloping shield volcanoes. They have broad summits and generally concave flanks. The base of Mauna Loa, a shield volcano on the island of Hawaii, is over 80 km in diameter. Shield volcanoes form by repeated eruptions from a common vent area of usually basaltic lavas low in SiO_2. These lavas have low viscosity and can flow for tens of kilometres. Examples of shield volcanoes in Canada include Mount Edziza, in central British Columbia, whose base is composed of several overlapping, large, basaltic

Figure 12.

Mount Baker is a typical stratovolcano, with steep sides and a pointed summit. Photograph by C.J. Hickson. GSC 2000-127

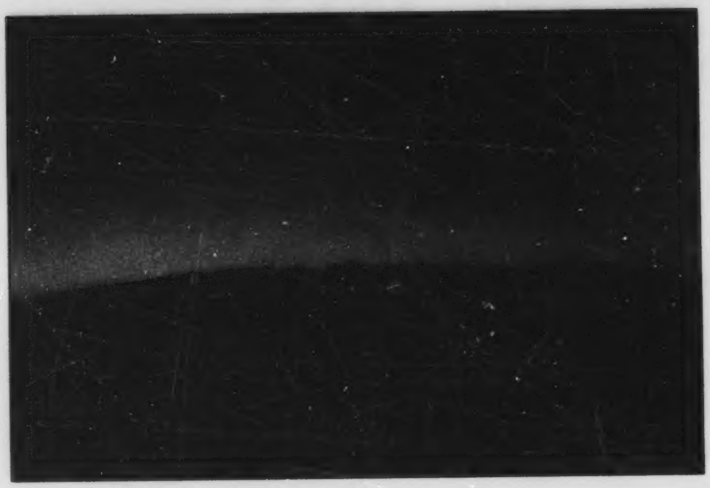


Figure 13.

The Illgachuz Range in western British Columbia. is a shield volcano several million years old. Photograph by C.J. Hickson. GSC 2000-128

shield volcanoes, and the Ilgachuz Range of central British Columbia (Fig. 13), which erupted special kinds of low-viscosity magmas to form a broad shield volcano whose youngest lava flows are several million years old.

Calderas form when large volumes of magma erupt all at once. The large hole left after the magma is ejected collapses, forming a circular to oval depression on the Earth's surface (Fig. 14). These holes are roughly similar in appearance to a crater formed by meteorite impact. Caldera-forming eruptions are the largest and most deadly types of volcanic eruptions. Fortunately, they are also very rare. Only one or two occur every several hundred thousand years. One of the most active regions for these eruptions is the western United States. Yellowstone Park is part of a 45 km diameter caldera formed 600 000 years ago. At least a few calderas may exist buried beneath the ice caps of volcanoes in Canada, including at Mount Silverthrone in southwestern British Columbia and at Mount Edziza in north-central British Columbia.

Other types of volcanoes are generally smaller, but mimic their larger cousins in shape if not in size. Cinder cones form around the vent where magma reaches the surface and are generally round to oval in plan view with a crater or depression at the top (Fig. 15). They are made of 'cinder', which is ash- to bomb-sized, solidified magma largely composed of vesicles (or solidified gas bubbles). Small, fist-sized, vesicular pieces of cinder can be referred to as 'scoria' (Fig. 9). As the magma reaches the surface, gases held in suspension within it are released rapidly — much like the bubbles that fizz out of a carbonated beverage the moment it is opened. Cinder cones are abundant in Western Canada (e.g. Wells Gray and Mount Edziza) and are easily accessible in some areas such as Atlin and Nazko, British Columbia.

Maars are volcanoes formed by the violent interaction of red-hot magma coming into contact with groundwater. The instantaneous conversion of the liquid groundwater to steam drives a violent explosion that forms a hollowed-out crater, similar to that made by a meteorite impact. The only

Figure 14.

The circular feature in the centre of the photograph is the summit caldera on Muana Loa volcano in Hawaii. Photograph courtesy of D. Peterson, United States Geological Survey.

Figure 15.

Eve cone is a young, well preserved cinder cone at Mount Edziza, British Columbia. Photograph by C.J. Hickson. GSC 2000-129

maar-like volcano known in Canada is found in Wells Gray Provincial Park, east-central British Columbia. Another type of volcano, unique to a relatively rare type of rock, is a diatreme. A diatreme is a complex and chaotic pipe of a special type of very low viscosity magma (such as kimberlite) filled with rock fragments. The eruption forces the column of magma and rock fragments to the surface from deep within the Earth's crust. Kimberlite diatremes or pipes have been found across Canada and the world and recently in the Northwest Territories. They are important economically because kimberlites are the world's main source of gem-quality diamonds. The kimberlite pipes found in the Northwest Territories and in Alberta may soon make Canada one of the world's major producers of gem-quality diamonds.

A distinctive type of volcano is produced when a volcanic eruption occurs beneath a glacier. This occurs most commonly today in Iceland and Antarctica. Both of these areas contain large glaciers (or ice caps) and active volcanoes. However, as recently as 10 000 years ago, much of Western Canada was also covered with thick sheets of ice. In fact, large parts of Canada have been repeatedly covered by thick glacial ice over the last few million years. As a result, Western Canada also has an abundance of volcanoes whose unusual, flat-topped (Fig. 16) or conical (Fig. 17) forms are the direct result of subglacial eruptions.

Dr. W.H. Mathews, a professor at the University of British Columbia, Vancouver, was one of the first people in the world to describe in detail these unique types of volcanoes while mapping in northern British Columbia in the 1940s. His work led him to recognize a sequence of interactions between a growing, subglacial volcano and the overlying ice that ultimately would produce the distinctive shapes of the subglacial volcanoes. He proposed that as a volcano erupted beneath a glacier, the heat from the volcano would immediately start to melt the overlying ice (Fig. 18, stage1). The resulting water quickly cooled the lava and produced forms called 'pillows'; aggregates of pillows form 'pillow lavas' (Fig. 19a, b). In places, the pillows broke up and rolled down the slopes of the submerged volcano, forming other types of volcanic deposits called 'pillow breccia' (Fig. 20), 'tuff breccia', and

Figure 16.

Hyalo Ridge, in Wells Gray Provincial Park, is a tuya, a typical, flat-topped, steep-sided, subglacial volcano. Photograph by C.J. Hickson. GSC 2000-130

Figure 17.

The 3 Caribou tuya, in north-central British Columbia, is a subglacial mound. Photograph by C.J. Hickson. GSC 2000-131

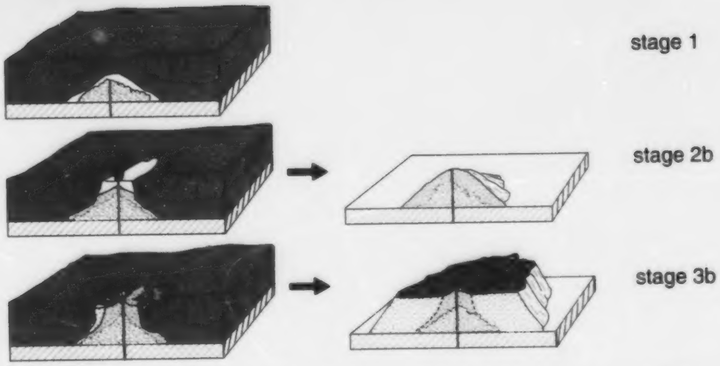


***Figure 18.** Stages in the development of subglacial volcanoes. The initial eruptions produce pillow lavas and hyaloclastite as the overlying glacial ice melts (stage 1). If the eruption stops before the volcano is higher in elevation than the surrounding water, a subglacial mound results (stage 2b). If the eruption continues, a subaerial lava flow may form a flat cap on the volcano, producing the classic tuya form (stage 3b).*

'hyaloclastite' (Fig. 21). Pillow lavas and other types of subaqueous (formed beneath water) volcanic deposits can be seen forming today in Hawaii where lava flows enter the crystal blue water of the Pacific Ocean.

If the volcano was not hot enough to melt through the overlying ice cap, a subglacial mound would be all that remained of the volcano once the ice had retreated (e.g. Fig. 17, 18, stage 2b). However, if the volcano was hot enough to melt through the overlying ice cap, more-normal-looking, subaerial (formed in the air) lava flows might flow out on top of the pillow lavas and cover them up. Once the glaciers had retreated or melted away, the steep-sided and in places flat-topped volcanoes would reappear with a unique shape as a result of their confinement within glacial ice (Fig. 18, stage 3b). Dr. Mathews referred to these flat-topped, subglacial volcanoes as 'tuyas' after Tuya Butte in northern British Columbia where he first studied these unique volcanic forms. Similar mountains in Iceland are called 'table mountains' because of their flat tops.

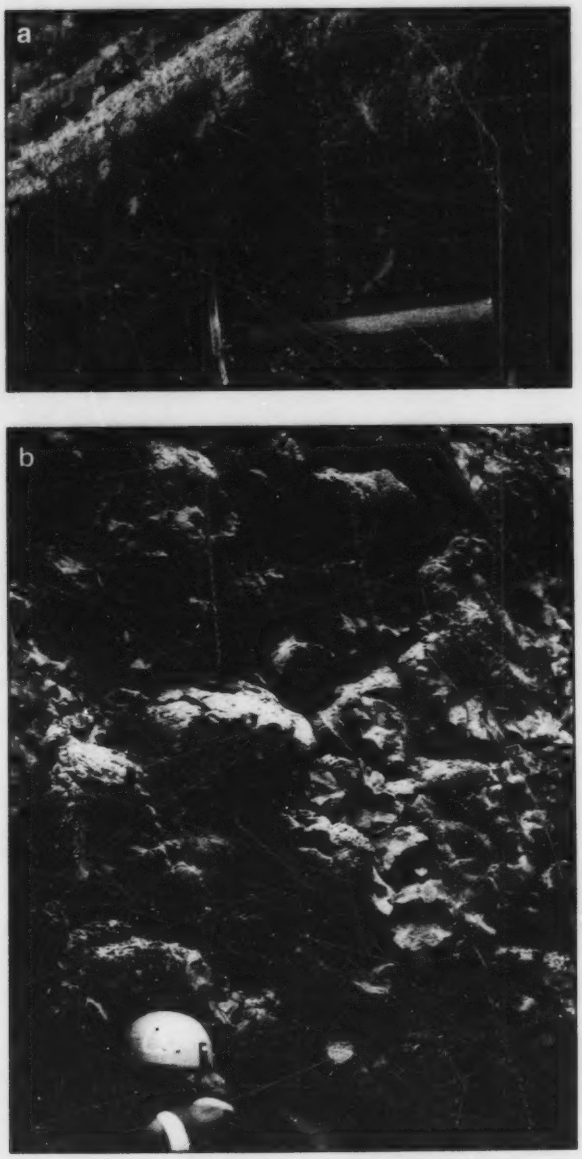


Figure 19. a) A single pillow, surrounded by hyaloclastite. Photograph by C.J. Hickson. GSC 2000-132 b) Stacks of pillow lavas commonly form where large lava flows enter a lake or an ocean. Photographs by C.J. Hickson. GSC 2000-133

VOLCANIC ERUPTIONS

Volcanoes are referred to as 'active', 'dormant', or 'extinct'. In general, an active volcano is one that has erupted within historic time. In regions where the historic record is very short (such as Canada), an active volcano is defined as one that has erupted within the last 200 years. In most cases, these

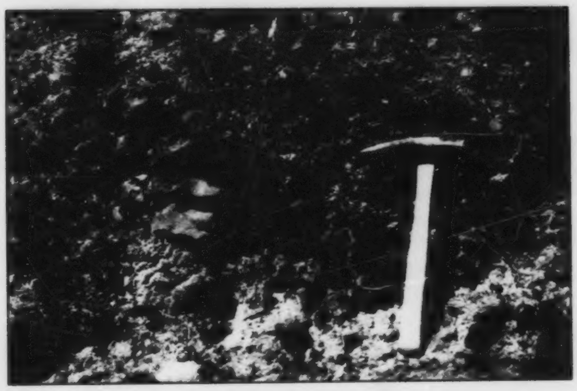


Figure 20. Fragments of pillow lava are commonly found with hyaloclastite. Together the hyaloclastite and pillow fragments are referred to as 'pillow breccia'. Photograph by C.J. Hickson. GSC 2000-134

Figure 21. Hyaloclastite is a glass-rich breccia that forms when volcanoes erupt under water or under ice. Most fragments are lapilli to ash sized. Photograph by C.J. Hickson. GSC 2000-135

volcanoes are expected to erupt again. A volcano that is actually erupting is said to be 'active' or 'in eruption'. A dormant volcano is one that has not erupted within historic time, but is expected to erupt again. An extinct volcano is not considered likely to erupt again.

When magma comes out of a volcano, either as lava flows or pyroclastic debris, the volcano is said to be 'erupting'. Volcanic eruptions can last from minutes to days. Commonly, a volcano will undergo several eruptions closely spaced in time. These closely spaced eruptions are part of the same eruptive episode or cycle. Such cycles can be passive (dome building or very small and local ejections of volcanic bombs referred to as 'fire fountaining') or explosive. For example, the most recent eruptive episode at Mount St. Helens (in southwestern Washington State; Fig. 2, 3) included numerous minor explosions (Fig. A-31), six major explosive eruptions, and passive extrusion of lava to form a dome above the volcano's vent. All this occurred over six years.

Mount St. Helens provides a good example of how a volcanic eruption can start and how the style of eruption can vary over time — even though the magma remains chemically similar. After over 100 years of dormancy, hundreds of small earthquakes occurred very closely spaced in time (an 'earthquake swarm'), starting 20 March 1980, heralding renewed activity at Mount St. Helens. By 27 March, small steam explosions had opened two small craters on top of the volcano. Steam explosions, some with small amounts of ash, continued intermittently until the catastrophic eruption began at 08:32 h on 18 May 1980. The gas-rich, dacitic magma exploded in an eruption that sent a column of tephra over 24 km into the atmosphere. Five more large explosive eruptions occurred through mid-October 1980. After that, as the volatile content of the dacitic magma decreased, activity became limited to the extrusion of lava and the formation of a lava dome over the vent.

Volcanic eruptions are commonly described by comparing the styles of eruption to the styles at well known volcanoes. Thus, Hawaiian-type eruptions are characterized by fire fountaining and the effusion of fluid lava flows. They are commonly not very hazardous. At the other extreme, Plinian-type eruptions, named for Pliny the Younger who first described them while watching the eruption of Mount Vesuvius in AD 79, are characterized by very large, violent, and very dangerous explosions. The 18 May 1980, eruption of Mount St. Helens was a Plinian eruption. In between these two extremes of danger and explosiveness are a variety of other types of eruptions with names such as 'surtseyan', 'vulcanian', 'strombolian', and 'pelean'. However, it is important to remember that an erupting volcano may exhibit more than one type of behaviour, and it is not uncommon for both passive and explosive phases to occur in a single eruptive cycle.

HAZARDS ASSOCIATED WITH CANADIAN VOLCANOES

Erupting volcanoes can generate many primary hazards including lava flows, pyroclastic flows, pyroclastic surges, volcanic bombs, ash clouds, landslides, debris flows, and clouds of poisonous gas (Fig. 22). The style of eruption and type of hazard generally depend largely on the size of the eruption, the composition of the erupting magma, and the environment in which the eruption occurs. For example, a small, subaerial, basaltic eruption in a remote part of British Columbia (e.g. close to Mount Edziza in northwestern British Columbia, Fig. 23) would constitute a minimal hazard compared with a large, explosive, dacitic eruption from Mount Meager, in southwestern British Columbia. However, even in remote areas such as Mount Edziza, an eruption can have a considerable impact if it occurs under a glacier-clad summit. Debris flows (lahars) or floods could form from the rapidly melting snow and ice. These floods, carrying large amounts of water and debris, might destroy salmon runs in the surrounding rivers and threaten local villages along these rivers.

The size of a volcanic eruption is quantified using a scale called the Volcanic Explosivity Index (VEI), which takes into account the volume of material erupted, the height of the eruption cloud, the duration of the main eruptive phase, and other parameters to assign a number from 0 to 8 on a linear scale. For example, the 18 May 1980 eruption of Mount St. Helens, which destroyed 632 km^2 of land, expelled 1.4 km^3 of magma, and produced an eruption column that rose to 24 km, was assigned a VEI of 5. On the other hand, the last large eruption from the Yellowstone caldera, which occurred 600 000 years ago and expelled over 1000 km^3 of magma, would be assigned a VEI of 7. However, the vast majority of volcanic eruptions have VEIs from 0 to 2.

Several secondary hazards are also associated with volcanoes. Secondary hazards are those that are not associated with an eruption, but rather result from the environment created by the volcano. They include mudflows, debris flows, landslides, ground water contamination, and surface water contamination (Fig. 22). All these hazards may persist at a volcano for decades after an eruption and even long after the volcano is considered extinct. Stratovolcanoes are particularly susceptible to debris flows because they are commonly very steep and are found in regions of rugged terrain. Heat from deep-seated cooling magma causes hydrothermal alteration of the overlying rocks. In this type of alteration, hot water attacks the volcanic rocks, turning them into clay. The presence of clay weakens the rocks and increases the risk of landslides.

Lava flows

Lava flows commonly accompany volcanic eruptions, especially basaltic eruptions. They are among the least hazardous processes associated with a volcanic eruption, although they can destroy any immobile object in their path, including houses, roads, and volcano observatories! Flows travel slowly, a few kilometres an hour to a fraction of a kilometre an hour, and people and animals can normally move easily out of the way. Lava flows, however, can cause many secondary hazards. They can start forest fires and will usually destroy any structures in their path. The expulsion of poisonous gases can also accompany the eruption of a lava flow, as happened in 1783 at the Laki volcano in Iceland. A lava flow can dam rivers, change their courses, kill resident fish, and even act as a barrier to migrating fish. This happened along the Nass River in northern British Columbia during the eruption of the Tseax cone. Of special concern in Western Canada is the possibility of lava erupting onto or beneath glacial ice. Such an event might cause large-scale melting of the ice and generate a catastrophic flood; such floods are referred to by their Icelandic name, 'jökulhlaups'.

Pyroclastic flows

Pyroclastic flows are dense avalanches of hot gas, hot ash, and blocks (tephra) that cascade down the slopes of the volcano during an eruption (Fig. 22). They are most commonly associated with explosive eruptions, and form when the towering column of ash rising above the volcano collapses. They also form when a less energetic eruption 'boils' over the edge of a crater or caldera or when a lava flow or dome on a steep slope disintegrates. Pyroclastic flows originating from column collapse can have considerable momentum and travel great distances. Speeds between 50 and 150 km/h have been measured and distances of 30 km are not unusual. Although extremely rare, the largest eruptions have produced pyroclastic flows that extend 100 km from the volcano. All pyroclastic flows are extremely destructive, destroying buildings, trees, or any objects that are in their path by impact, burial or fire. People caught in a pyroclastic flow have little chance of survival. Pyroclastic-flow deposits are not abundant at dormant volcanoes in Canada, but have been found at Mount Meager, Hoodoo Mountain, and Mount Edziza (Fig. 23; *see* Appendix A).

Pyroclastic surges

Pyroclastic surges are dense clouds of hot gas and rock debris, that are generated when water and hot magma interact. They are more violent and travel much faster than pyroclastic flows; surges have been clocked at over 360 km/h. They are extremely destructive because of their density and high speeds. People and structures in their path have little hope for survival. White Horse Bluffs, in British Columbia, was built from successive surge deposits, but in general, these deposits are difficult to preserve and what little evidence may exist is easily eroded away.

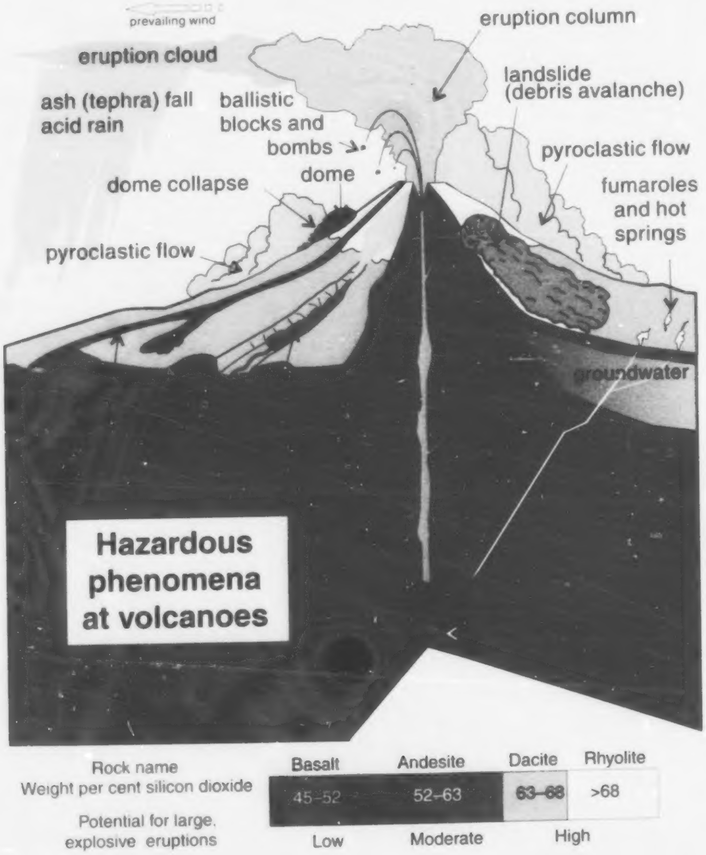

prevailing wind
eruption cloud
ash (tephra) fall
acid rain
ballistic
blocks and
bombs
eruption column
landslide
(debris avalanche)
pyroclastic flow
dome
dome collapse
fumaroles
and hot
springs
pyroclastic flow
groundwater
Hazardous
phenomena
at volcanoes
Rock name
Weight per cent silicon dioxide
Potential for large,
explosive eruptions
Basalt
Andesite
Dacite
Rhyolite
45–52
52–63
63–68
>68
Low
Moderate
High

Figure 22.

Schematic figure illustrating hazards associated with different types of volcanic eruptions. Most commonly, not all hazards shown will accompany a single eruption from a volcano, although a single eruption can produce more than one type of hazard at the same time. The rock classification at the bottom emphasizes that magmas with greater amounts of SiO_2 *(weight per cent) are more likely to produce large, explosive eruptions* (modified from *Myers et al., 1998).*

Tephra and ballistic projectiles

Tephra is finely comminuted (broken) volcanic rock and accompanies nearly all explosive volcanic eruptions. In very energetic, explosive eruptions, tephra is carried upward into the upper atmosphere and the finest tephra (ash) can be carried by the jet stream for hundreds and thousands of kilometres. Ballistic projectiles (e.g. volcanic blocks and bombs) are the larger blocks of desegregated and comminuted magma that are most commonly thrown short distances from the vent (Fig. 22), but that are know to have been thrown

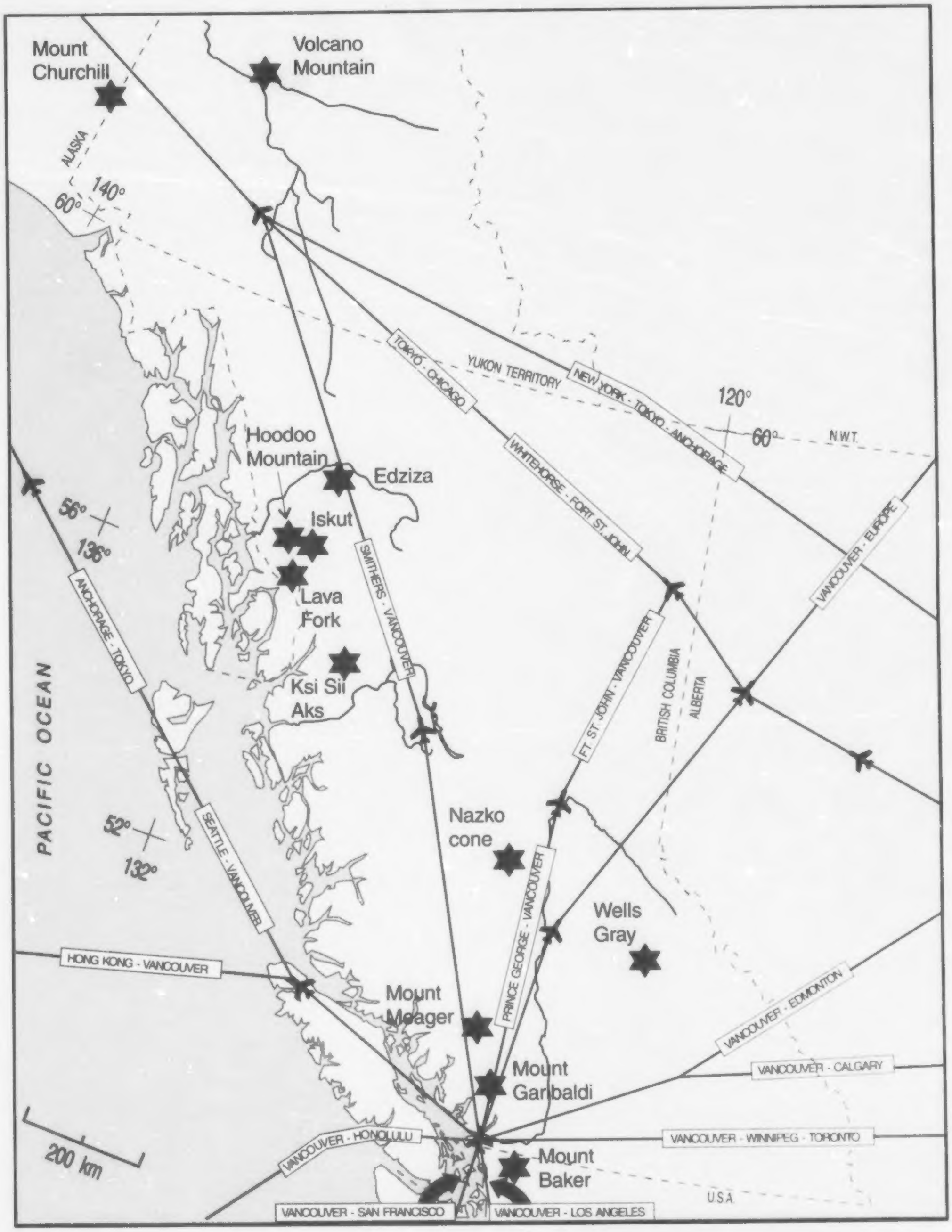

Mount Churchill
Volcano Mountain
ALASKA
140°
60°
TOKYO - CHICAGO
YUKON TERRITORY
NEW YORK - TOKYO - ANCHORAGE
120°
60°
N.W.T.
Hoodoo Mountain
Edziza
Iskut
WHITEHORSE - FORT ST. JOHN
56°
136°
SMITHERS - VANCOUVER
VANCOUVER - EUROPE
Lava Fork
ANCHORAGE - TOKYO
PACIFIC OCEAN
Ksi Sii Aks
FT. ST. JOHN - VANCOUVER
BRITISH COLUMBIA
ALBERTA
Nazko cone
52°
132°
SEATTLE - VANCOUVER
PRINCE GEORGE - VANCOUVER
Wells Gray
HONG KONG - VANCOUVER
Mount Meager
VANCOUVER - EDMONTON
Mount Garibaldi
VANCOUVER - CALGARY
200 km
VANCOUVER - HONOLULU
VANCOUVER - WINNIPEG - TORONTO
Mount Baker
USA
VANCOUVER - SAN FRANCISCO
VANCOUVER - LOS ANGELES

Figure 23. *Diagram showing major air routes over Western Canada. The volcanoes marked with stars are described in more detail in Appendix A.*

several kilometres from the vent. In less energetic eruptions, most of the tephra falls within a few kilometres of the vent and the accompanying ballistic projectiles, a correspondingly shorter distance.

Because it is often widely dispersed, volcanic ash may be a significant health hazard and create economic problems over a wide area. Ash can pollute water supplies and disrupt transportation; thick accumulations of heavy ash can cause buildings or other structures to collapse. Inhaled ash can aggravate respiratory conditions such as asthma and bronchitis. Coarser particles can lodge in the nose (causing irritation) and the eyes (causing corneal abrasions). Silicosis has been attributed to long-term exposure to volcanic ash. Ash, however, is rarely a direct cause of death unless the fallout is so severe that suffocation results.

Ash can damage mechanical and electrical equipment. It is abrasive and, at great distances from the volcano, fine enough to work its way into bearings or other moving parts of equipment and machinery, causing damage or mechanical failure. Computer equipment is particularly sensitive to this type of damage. Electrical power can be disrupted because transformers conduct heat poorly and will overheat and explode when covered by only a few millimetres of ash.

Ash can affect aircraft flying at high altitudes. The adverse impact of ash on high-performance jet engines was well publicized for the first time in 1982 when a Boeing 747 jet on a flight over Indonesia encountered ash from Galunggung volcano. Although the plane was severely damaged and all four engines failed, it was able to make a successful emergency landing in Jakarta. Since then, the number of plane–volcanic ash incidents has grown and the adverse effects of ash have been studied more carefully. Ash sucked into high-performance jet engines abrades the outer parts of the engine and can melt inside at the high engine operating temperatures. The abrasion severely damages the engine fan blades and the melting ash disrupts the mixing of air and fuel in the combustion chamber, leading to engine shutdown. The ash also abrades the exterior of the aircraft, 'frosting' cockpit windows and landing lights. Ingested ash can badly damage aircraft navigation and hydraulic systems. The threat from airborne ash in Canadian airways constitutes the most important short-term impact of volcanoes on the Canadian public (Fig. 23).

Ash also has deleterious effect on fish because it abrades their gills. Studies of steelhead and salmon in streams west of Mount St. Helens showed that fish populations suffered severely after the 1980 eruptions. There were many other lethal, secondary effects, such as the loss of fish-spawning habitat and riparian cover. Thick accumulations of ash and mudflows can kill the riparian cover (the normal trees and bushes that grow along steams and rivers), which keeps stream waters cool and liveable for fish.

Thick accumulations of ash can significantly affect forestry and agriculture. Ash can rip the leaves from trees and bury small plants. Ash-covered trees and crops are difficult to work because of the potential damage to harvesting and cultivation equipment and vehicles caused by the ash. Wind and moving equipment also redistribute the ash, prolonging the problems. However, the long-term impact of ash can actually be beneficial, as ash enriches the soil and, as with tephra from the 1980 Mount St. Helens eruption, can act as a mulch, increasing soil moisture and promoting growth.

Edifice or sector collapse and landslides and debris avalanches

This hazard, known by several names, involves the collapse of part of the volcano. If collapse is associated with an eruption, it can result in a larger and potentially more explosive eruption than might otherwise be expected. Such was the case in the 18 May 1980 eruption of Mount St. Helens, where catastrophic depressurization of the volcano caused by a massive landslide resulted in an eruption of significantly different character than what had been predicted. As magma is pushed into a volcano from the magma chamber below via the magma conduit, the volcano inflates, much like blowing up a balloon (Fig. 22). During inflation, the volcano's slopes become oversteepened and unstable. At this stage, ground shaking associated with stream venting and magma ascent can trigger a collapse. The collapse of a large part of the edifice is termed a 'sector collapse'.

Landslides and debris avalanches originating on volcanoes also constitute a significant secondary hazard unrelated to volcanic eruptions. As a landslide moves downslope, the large blocks break into smaller and smaller pieces, which, when combined with water from melting snow or ice, can transform the landslide into a debris avalanche, carrying it farther than a similar-sized rock avalanche. The slopes of volcanoes, steepened by explosions or later erosion, are often weakened by hydrothermal alteration, making them susceptible to failure long after the volcano becomes dormant.

Lahars

'Lahar' is an Indonesian word for debris flow . These slurries of water and rock particles behave like wet concrete and are associated with volcanic eruptions. They consist of particles of many different sizes, ranging from flour-sized particles to blocks as large as houses. Lahars are extremely destructive; they are, however, topographically controlled and usually follow river valleys where they are confined to valley bottoms. They are most common and most voluminous as a result of explosive eruptions on snow-clad volcanoes. Pyroclastic flows can instantaneously melt ice and snow, creating large-volume lahars.

Debris flows can be a significant secondary volcanic hazard. They can occur days, weeks, or years after an eruption. Explosive eruptions and, to a lesser extent, effusive eruptions can denude areas around a volcano and disrupt the drainage pattern, which can lead to long-term flooding problems around the volcano. In addition, vast areas around the volcano may be covered by loose, unconsolidated tephra, which is easily mobilized and can be washed away by heavy rainfall, forming mudflows and debris flows.

Acid rain and gases

Volcanoes produce large quantities of gases, mostly water (H_2O), but also significant amounts of carbon dioxide (CO_2), hydrogen sulphide (H_2S), sulphur dioxide (SO_2), and hydrofluoric acid (HF) as well as some chlorine (Cl) and nitrogen (N) compounds. Each of these gases has caused loss of life and damage. Sulphur dioxide vented from the Laki volcano in Iceland in 1783 damaged crops and killed livestock and people. Survivors faced starvation and many more died as a result. Sulphur dioxide causes other problems as well. Exposure to this gas can lead to more numerous incidents of acute asthma and bronchitis. When SO_2 reacts with water in the atmosphere, it forms sulphuric acid and acid rain. Acid rain formed from volcanic clouds can damage crops and, in Hawaii, the increased acidity of water collected in cisterns leaches heavy metals into drinking water. The long-term health affects of this problem are not completely known. In Iceland, fluorine killed and disfigured livestock after the 1845 and 1970 Hekla eruptions. When CO_2 gas, which is heavier than air, is released by a volcano, it collects in low areas; concentrations can become high enough to be deadly. The only known fatalities at a Canadian volcano are reported to have been from poisonous gases during the eruption of the Tseax cone.

Groundwater

A secondary hazard at volcanoes is contamination of groundwater and surface water. By virtue of the composition and physical attributes of their component rocks, volcanoes are much more susceptible to weathering than many other types of rocks. Weathering leads to increased particulate matter in streams draining the edifices and higher concentration of elements easily leached by percolating groundwater. Thermal springs are often associated with volcanoes. The water in some of these springs is acidic. Hot, acid waters will enhance the breakdown (or leaching) of the rock and will often carry dissolved metals. Because of this, streams and rivers draining volcanic areas may have a significantly different trace-element chemistry than nearby streams draining other regions underlain by more chemically stable rock types.

REDUCING THE RISK FROM VOLCANOES

Erupting volcanoes only become a risk when they affect something of value, i.e. by causing injury or death to people or damage to property or resources. Risk is an indication of the possible impact of a particular hazard, a lava flow or pyroclastic flow for example. It is usually assessed on the basis of the number of human lives that may be lost because of the hazardous event. In 1000 years of record keeping, volcanic eruptions have killed less than 300 000 people. Wars and other hazardous events have taken many millions of human lives. Volcanic eruptions are far more likely to affect the economic base of a region than to kill people. The economy of an area can be affected by the destruction of either resources or infrastructure. Even eruptions in remote areas can adversely affect industries such as logging and fisheries.

Assessments of volcanic hazards made for individual volcanoes provide a basis for understanding the type of activity (hazards) that might be expected at that volcano in the future, given its previous history. From these studies and other factors, such as population density, risk can be estimated. These assessments, carried out for the American portion of the volcanoes of the Cascade Mountains by the United States Geological Survey, provide a basis for long-range planning in areas likely to be affected by a volcanic event. They are, however, limited by the level of knowledge about an individual volcano. The risk may be significantly underestimated because of incomplete knowledge about the eruptive history of an area. The record for most Canadian volcanoes is too poorly known to make accurate hazard assessments.

Growing populations increase the risk posed by volcanoes both in Canada and abroad. For example, neither Mount Pinatubo, in the Philippines, nor Nevado del Ruiz, in Columbia, were considered high-risk volcanoes. Loss of life from the 1991 eruption of Mount Pinatubo was minimized because excellent monitoring provided enough time to evacuate people from critical areas. Nevado del Ruiz was monitored, but warnings failed to reach the people. When it erupted on 13 November 1985, 25 000 people were killed by mudflows — the greatest volcanic disaster since the 1902 eruption of Mount Pelé in the Caribbean killed 28 000 people. A poignant point brought out in later studies of the Nevado del Ruiz eruption was the observation that in 1845, a similar mudflow wiped out 1400 people, virtually the town's entire population. In 1985, 30 000 people lived in the same area. Twenty times more people died in the 1985 eruption than in the earlier event. In a similar vein, in the 1814 eruption of the Philippine volcano of Mayon, pyroclastic flows killed 1200 people, most of the inhabitants living in the area — 800 000 people now live in the same region.

Lava flows

Lava flows pose little threat or significant risk unless the eruption occurs in winter or beneath or near glacial ice. If lava flows over significant amounts of snow, it may lead to some melting, producing debris flows that extend past the end of the lava flow. In addition, water entering a basaltic vent can produce explosions much larger than those normally associated with a basaltic eruption. Consequently, the presence of water, snow, or ice will increase the risk of significant impact from an eruption. Subglacial volcanism has also produced large, destructive jökulhlaups.

Ballistic projectiles and other flowage phenomena

Pyroclastic flows, surges, lava flows, and ballistic projectiles constitute a significant risk only to communities and resources close to the volcano. As a rule of thumb, studies suggest that areas within 50 km of an erupting volcano may be severely affected by these and other volcanic hazards. Beyond this limit, the main hazards (and thus risk) are from tephra and debris flows, the effects of which can be localized because of prevailing wind patterns and topography.

Tephra

Tephra poses the greatest risk to Western Canada. In the past 12 000 years, volcanoes in the Cascade magmatic arc have erupted over 200 times. Several of these eruptions deposited significant quantities of tephra in southern British Columbia, most of it fine, ash-sized particles in a layer a few millimetres in depth (Fig. 6). Alaskan volcanoes in the Aleutians have erupted even more often, dusting and blanketing parts of northern Canada with ash (e.g. White River Ash, Fig. 6 in pocket). During the 1992 eruption of Mount Spurr in Alaska (Fig. 6 in pocket), ash blown southeastward over Canada by upper-atmospheric winds crossed over the Yukon Territory into British Columbia and Alberta, then blew over Ontario and Quebec before dispersing over the North Atlantic (Fig. 24). Despite the proximity of active and erupting volcanoes in the United States, significant accumulations of tephra (over 10 cm) are very rare. We can, however, expect small amounts of ash much more frequently. The 18 May 1980 eruption of Mount St. Helens deposited a 1 mm thick layer of ash from southeastern British Columbia to Manitoba. During its 1992 eruptions, Mount Spurr deposited enough ash in the Yukon Territory to close the Alaska Highway for a few hours due to reduced visibility.

Landslides, sector collapse, and debris avalanches

This hazard is a risk only close to a volcano. Large rock failures (or landslides) are a hazard at many of Canada's volcanic centres, even those long extinct. Many of Canada's volcanoes are in extremely rugged regions of high relief and are underlain by unstable rocks. These conditions have already led to many landslides. One of these, at Mount Meager in southern British Columbia, killed four people. Such rock failures can generate secondary hazards when they temporarily dam local rivers, which may later lead to dam failure and catastrophic flooding.

Lahars and debris flows

This hazard poses a risk to some communities situated along the banks of rivers or streams draining directly from a volcano. During one Mount Meager eruption, 2350 radiocarbon years ago, ash was carried as far as Alberta and pyroclastic flows extended 7 km down the Lillooet River. Upstream of the eruption, volcanic debris dammed the Lillooet River, forming a lake. Catastrophic failure of this dam created a flood. The flood waters carried 10 m blocks of volcanic debris several kilometres downstream and inundated large parts of the valley downstream from the eruption. Eruptions almost anywhere along the valley corridor between Vancouver and Pemberton may generate lahars that will cut rail and road links, as do the nonvolcanically induced landslides and debris flows that are common in that area.

MONITORING CANADA'S VOLCANOES

In Canada, even though some volcanoes pose a significant threat to local communities and any sizable eruption would affect the economy of Western Canada, research is limited. Because no large eruptions have occurred in Canada in the last few hundred years, volcano monitoring is not a priority. Existing seismic networks could be used to identify activity at a Canadian volcano should one reawaken, but in many areas, this network will alert authorities only if the impending eruption is potentially large. Scientists at the Geological Survey of Canada and at Canadian universities have carefully studied the past behaviour of a number of Canadian volcanoes. These studies are critical to understanding the future behaviour of a volcano. Additionally, the threat from volcanoes outside of Canada seems much greater than the threat from Canadian volcanoes. Volcanoes in Alaska, Washington, Oregon, and California are active more frequently than those in Canada and are carefully monitored by the United States Geological Survey. A plan is in place to alert Canadian authorities of eruptions that may send ash into Canada.

Monitoring volcanoes has become more refined in the past few decades, in large part because of the scientific information gained from the eruptions of Mount St. Helens in the 1980s, but it is still fraught with difficulties. As noted earlier, at Mount Pinatubo, Philippines, monitoring precursory activity provided enough warning of the eruption for a timely evacuation of inhabitants. Geologically well studied volcanoes, monitored using several types of instruments, provide the best opportunity for predicting and forecasting future events. Large, explosive eruptions are usually preceded by precursory activity. If monitoring equipment is in place, it will likely provide some warning of an impending eruption. Basaltic eruptions, especially effusive ones, may have little or no precursory activity associated with the rise of magma. In addition to monitoring, geological mapping and dating of volcanic deposits are two basic, key factors for understanding and predicting how a volcano will behave the next time it erupts.

VOLCANOES ON THE INTERNET

The Internet is a valuable tool and aide for people interested in volcanoes and volcanic eruptions. Many governments and groups of scientists maintain Web pages presenting information about well studied volcanoes. Several volcano observatory home pages are updated with current information about volcanic activity around the world, including centres at the University of North Dakota, the Cascade Volcano Observatory, the Alaska Volcano Observatory, the Hawaii Volcano Observatory, the Monteserrat Volcano Observatory and the Nordic Institute (in Iceland).

ACKNOWLEDGMENTS

This contribution would not have been possible without the help of many people. The authors would especially like to thank Cecilia Porto (figures and photographs) and Bev

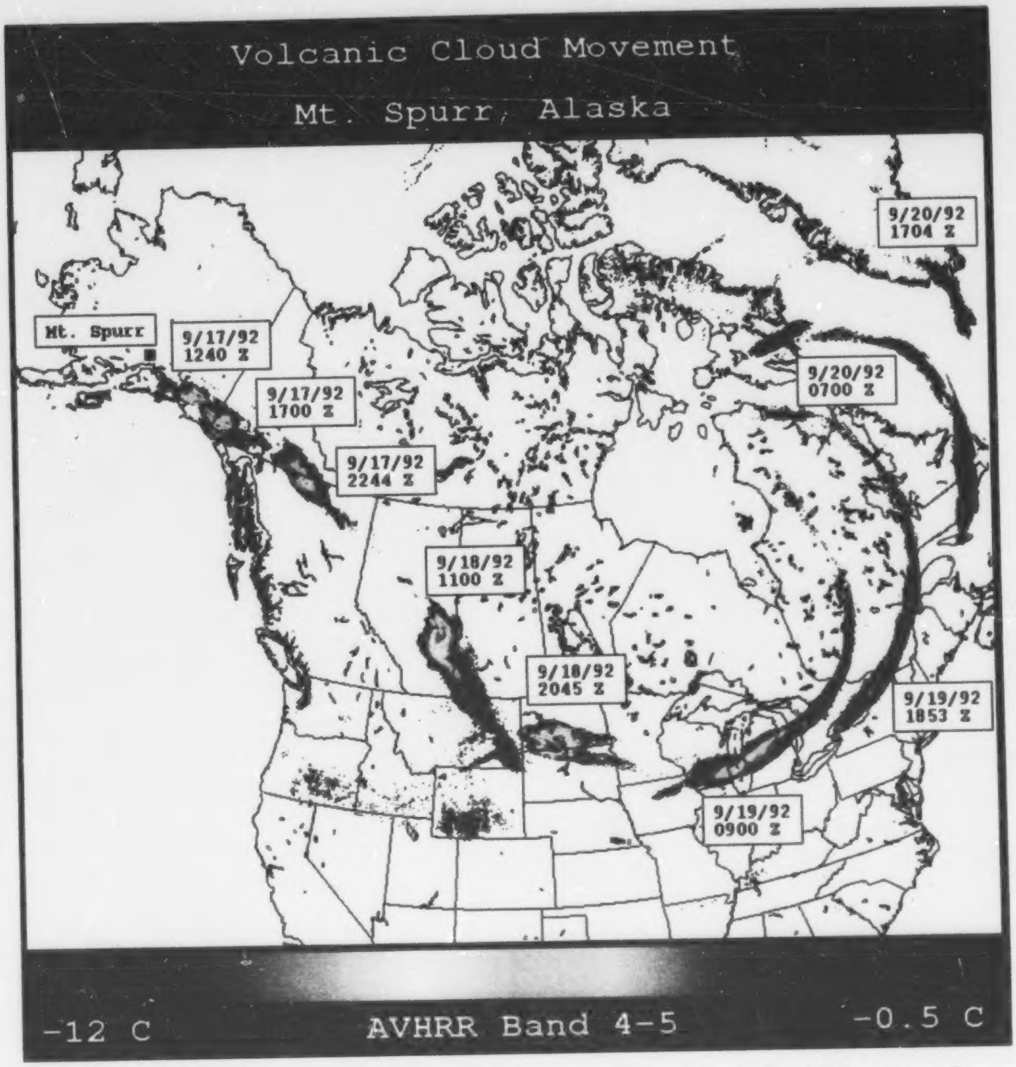

Volcanic Cloud Movement
Mt. Spurr, Alaska
Mt. Spurr
9/17/92
1240 Z
9/17/92
1700 Z
9/17/92
2244 Z
9/18/92
1100 Z
9/18/92
2045 Z
9/19/92
0900 Z
9/19/92
1853 Z
9/20/92
0700 Z
9/20/92
1704 Z
-12 C
AVHRR Band 4-5
-0.5 C

Figure 24. *Composite satellite radar image of tephra from the 17 September 1992 eruption of Mount Spurr in Alaska. The satellite radar tracked the ash cloud all the way across Canada to Greenland. (courtesy of D. Schneider, Alaska Volcano Observatory, United States Geological Survey)*

Vanlier (text editing) for their hard work. Steve Williams was tremendously helpful with graphics and is responsible for the GIS work that produced Figure 6. Crystal Huscroft assembled the background material for the ash-plume distributions. Her help is sincerely appreciated. We also thank Greg Brooks and Glenn Woodsworth for editorial comments on the text and figures. We thank Lionel Jackson for providing photographs of the White River Ash and Volcano Mountain and Murray Journeay for the Landsat image of the Garibaldi area.

REFERENCES

Barbeau, M.
1935: Canadian Geographical Journal; v. X, no. 5, p. 215–225.

Krafft, M.
1993: Volcanoes, Fire from the Earth; Thames and Hudson, London, United Kingdom

Myers, B., Brantley, S.T., Stauffer, P., and Hendley, J.W.
1998: What are volcano hazards?; United States Geological Survey Fact Sheet-002-97, revised June 1998.

Souther, J.G.
1992: The Late Cenozoic Mount Edziza Volcanic Complex, British Columbia; Geological Survey of Canada, Memoir 420, 320 p.

SUGGESTED READING

Chester, D.
1993: Volcanoes and Society; Edward Arnold, London, United Kingdom, 351 p.
Decker, R. and Decker, B.
1989: Volcanoes; W.H. Freeman and Company, New York, New York, 285 p.
Fisher, R.V., Heiken, G., and Hulen, J.B.
1997: Volcanoes: Crucibles of Change; Princeton University Press, Princeton, New Jersey, 317 p.
Francis, P.
1993: Volcanoes; Oxford University Press Inc., New York, New York, 443 p.
Harris, S.L.
1992: Fire Mountains of the West; Mountain Press Publishing, Missoula, Montana, 379 p.
Scarth, A.
1994: Volcanoes: an Introduction; Texas A&M University Press, 273 p.
Wood, C.A. and Kienle, J.
1990: Volcanoes of North America; Cambridge University Press, Cambridge, United Kingdom, 354 p.

APPENDIX A

Volcanoes in Canada and the United States

Mount Baker (northwestern Washington State, United States)
Location: 48.79°N, 121.82°W
Volcanic features: Stratovolcano, lava flows, pyroclastic flows, hydrothermal vents
Oldest known eruption: 900 000 years ago
Most recent known eruption: AD 1843 (tephra)
Summit elevation: 3285 m
Base elevation: 1650 m
Estimated volume erupted: 72 km^3

Mount Baker is a large, active, andesitic stratovolcano, rising to an elevation of 3285 m in the Cascade Range of northern Washington State (Fig. A-1). Its summit is approximately 100 km southeast of downtown Vancouver, British Columbia, 30 km south of the Canada–United States border, and 50 km due south of Chilliwack, British Columbia. It is visible on clear days from high points throughout this part of southwestern British Columbia.

Mount Baker is the largest volcanic complex in the northern part of the Cascade volcanic arc. Its volume is estimated at 72 km^3 and it supports one of the largest geothermal fields in the Cascade Range. The geothermal field found near the summit in the 450 m wide Sherman Crater periodically produces clouds of steam from hot springs and fumaroles (Fig. A-2).

Remnants of two older volcanoes underlie the existing cone. The oldest is a caldera complex measuring 5 km by 8 km. This caldera, referred to as 'Kulshan caldera', is comparable in size to Crater Lake. Overlying it is an approximately 400 000-year-old dacitic to basaltic volcano, Black Buttes, that has been largely destroyed by erosion. The existing cone was formed before the last major glaciation in North America (25 000–10 000 years ago) and consists of interlayered, andesitic lava flows and breccia. Holocene activity (younger than 10 000 years old) at Mount Baker included pyroclastic flows, lava flows, and the eruption of tephra from Sherman Crater. Tephra, as well as at least one lava flow, was also expelled from a vent near the south base of

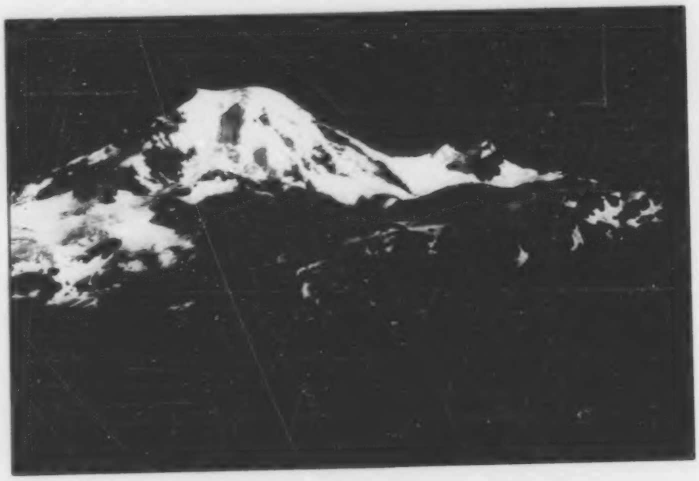


Figure A-1.

Mount Baker forms a prominent landmark visible from much of southwestern British Columbia. The peak of the 3285 m high stratovolcano is often shrouded in clouds, but on a clear day, steam can be seen rising above Sherman Crater, which is just out of view to the right of the summit in this photograph. Photograph courtesy of P. Hickson. GSC 2000-136

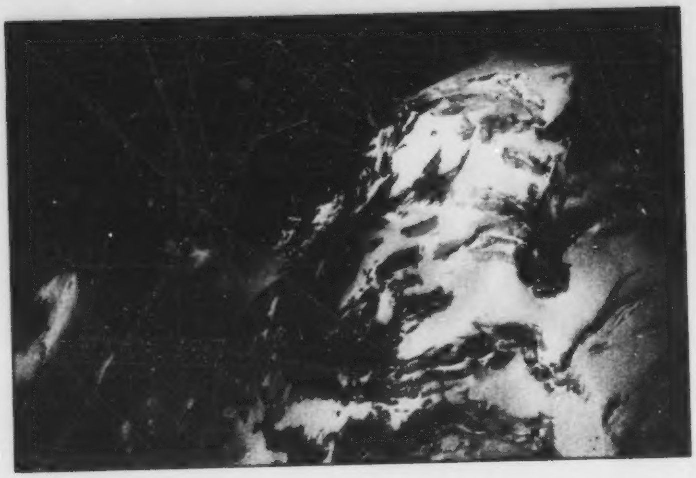


Figure A-2.

Snow shrouds the summit and much of the flanks of Mount Baker, but here, in the 450 m wide Sherman Crater, steam has melted through the ice, forming the 10 m wide hole along the west side of the crater. Steam also emanates from the steep back wall of the crater that leads to the summit. Photograph by C.J. Hickson. GSC 2000-137

the mountain at Schriebers Meadow. Increased hydrothermal activity, starting in March 1975 and tapering off in 1977, prompted a number of scientific studies to investigate the reason for the increased steam emissions. The United States Geological Survey is currently carrying out detailed geological investigations.

Mount Baker's next major eruption is most likely to send a column of ash several kilometres into the atmosphere and erupt viscous lava flows. Neither of these events constitute a significant hazard. However, because the mountain is capped by an estimated 2 km³ of glacial ice and snow, even minor eruptions could generate significant mudflows and floods. Depending on which valley these floods descended, cities in the eastern Fraser River valley (such as low-lying parts of Abbotsford) could be flooded. In addition, because the volcano is almost directly under flight paths into and out of Vancouver, British Columbia, even minor eruptions of ash constitute a significant hazard to aviation in southwestern Canada.

Larger, more devastating eruptions are known from Mount Baker's past. If the style of future eruptions at Mount Baker were to change from the smaller eruptions that have occurred over the past few hundred years, any future eruption could have a far more significant impact on southwestern British Columbia. An emergency preparedness exercise in January 1997 tested the ability of government and inhabitants of the Fraser River valley to cope with a significant volcanic event resulting from an eruption of Mount Baker. The exercise, referred to as 'Thunderbird 3', was run by the British Columbia provincial government in collaboration with other levels of government. Emergency preparedness personnel in communities in the Fraser River valley were asked to cope with significant accumulations of ash and with flooding of many low-lying parts of the valley. Exercises such as Thunderbird 3 are crucial for developing volcanic-hazard safety awareness and action plans in communities likely to be affected by future volcanic eruptions.

Garibaldi volcano (southwestern British Columbia)
Location: 49.83°N, 123.00°W
Volcanic features: Stratovolcano, aa lava flows, cinder cones, subglacial deposits, ice-dammed lava flows
Oldest known eruption: 300 000 years ago
Most recent known eruption: ~10 000 years ago
Summit elevation: 2678 m
Base elevation: 2130 m
Estimated volume erupted: >6.5 km³

Garibaldi volcano is an eroded, dacitic stratovolcano in southwestern British Columbia; like Mount Baker to the south, it is part of the Cascade volcanic arc. Garibaldi volcano, 80 km due north of Vancouver, is made up of Mount Garibaldi, Atwell Peak, and Dalton Dome (Fig. A-3, A-4). This Pleistocene volcanic centre is a part of a volcanic field that contains some 13 vents in an area 30 km long by 15 km wide, much of which is in Garibaldi Provincial Park. The eruptive history of Garibaldi volcano involves an initial period of volcanism (200 000–300 000 years ago) followed by a period of quiescence. Renewed activity in the last 50 000 years has rebuilt the edifice in a series of violent eruptions, similar in character to those that issued from Mount Pelée in 1902 and destroyed the town of St. Pierre on the island of Martinique. As successive pyroclastic flows travelled down the mountain's gentle slopes and were deposited, the volcano took on a broad, conical form. Part of the volcano's southwest flank built out onto thick glacial ice filling the Squamish River valley (Fig. A-3). Subsequent rapid melting of the ice at the close of the last ice age removed support from the western part of the cone. The volcano collapsed, producing much of the existing rugged topography of Mount Garibaldi and Atwell Peak. This catastrophic failure left a scarp on which is exposed the internal structure of the volcano and a debris fan with an estimated volume of 150 000 000 m³ at the foot of the mountain north of the town of Squamish. As valley glaciers retreated, two lava flows erupted from Clinker Peak, immediately north of Garibaldi volcano (Fig. A-3). The northernmost Rubble Creek flow was partly confined by a wall of ice, resulting in a lava flow over 244 m (800 ft.) thick.

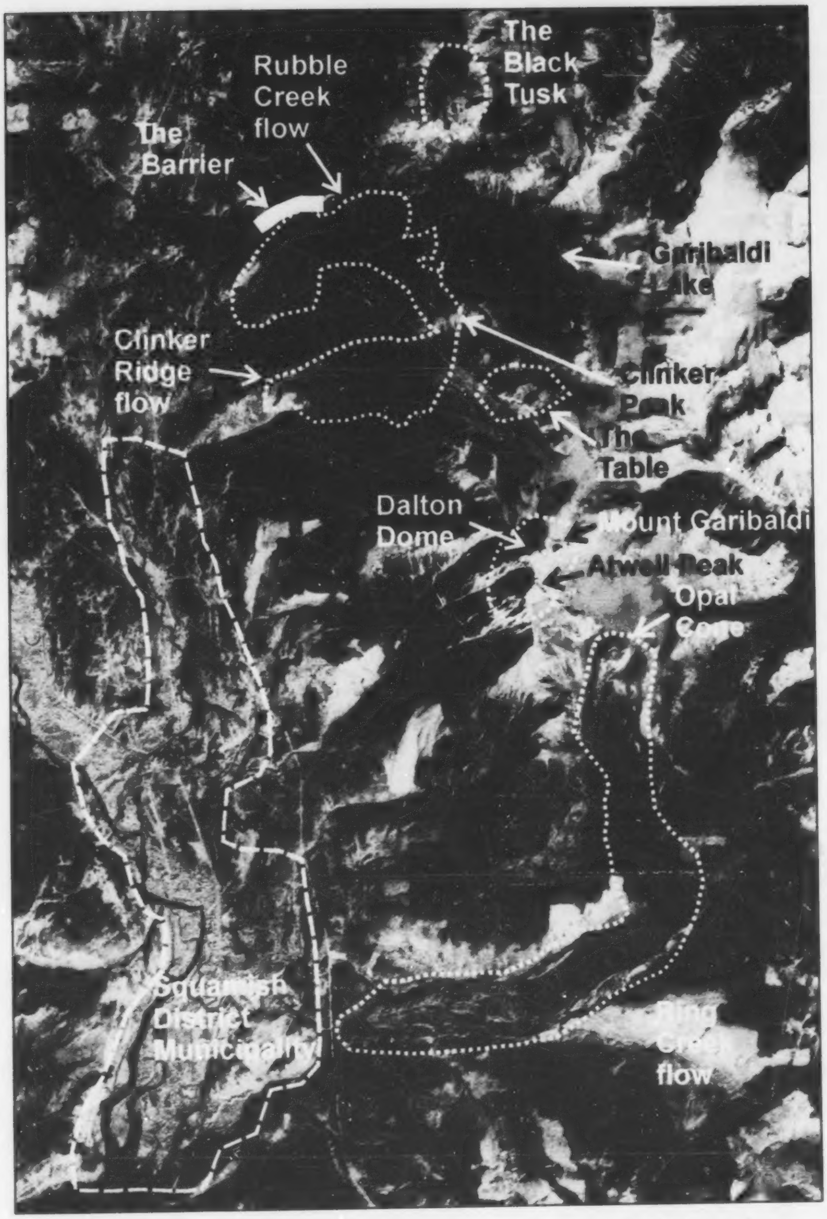

The Black Tusk
Rubble Creek flow
The Barrier
Garibaldi Lake
Clinker Ridge flow
Clinker Peak
The Table
Dalton Dome
Mount Garibaldi
Atwell Peak
Opal Cone
Squamish District Municipality
Ring Creek flow

Figure A-3. *This Landsat image of the area surrounding Garibaldi volcano shows several different volcanic features including lava flows (Ring Creek, Rubble Creek, Clinker Ridge), a volcanic neck (The Black Tusk), and a partly eroded stratovolcano (Mount Garibaldi, Atwell Peak, Dalton Dome). Also visible is The Barrier, which has been the source of several catastrophic rock avalanches. The approximate outline of the Squamish District Municipality is also shown for reference.*

The steep, northern edge of the Rubble Creek flow partly collapsed several times, most recently in 1855–1856 (Fig. A-5). The village of Garibaldi was abandoned because of the danger of future collapses. The most recent period of activity occurred shortly after the disappearance of the glacial ice filling the valley, 10 700 to 9300 radiocarbon years ago, and ended with the eruption the Ring Creek lava flow from Opal cone, on Garibaldi's southeastern flank (Fig. A-3). The Ring Creek flow is very unusual. It is 15 km long — a length usually only attained by fluid basalt flows, except that the Ring Creek flow is dacite.

Renewed volcanism in the Garibaldi area would pose a serious threat to the local communities of Whistler and Squamish. Although no Plinian-style eruptions are known, even Pelean-type eruptions could produce large quantities of ash that could significantly affect these nearby communities. Ash columns could rise to several hundred metres above the volcano whose close proximity to Vancouver would make this a hazard for air traffic. The danger from lava flows would be low to moderate because the nature of the lavas would prevent them from travelling far from their source, even though the Ring Creek lava flow ends only 6 km from Squamish. Melting of remnant glacial ice capping the Mount Garibaldi area could produce floods, lahars, or debris flows that might endanger small communities including Brackendale. Highway 99, which links Whistler and Squamish with Vancouver, is already plagued by landslides and debris flows from the precipitous Coast Mountains. An eruption producing floods could destroy segments of the highway. Flooding and debris flows could also have serious consequences for the salmon fishery on the Squamish, Cheakamus, and Mamquam rivers. In addition, explosive eruptions and the accompanying ash could cause short- and long-term water-supply problems for Vancouver and much of the lower mainland. The catchment area for the Greater Vancouver watershed is downwind from the Garibaldi area. Air-fall material could also have a deleterious effect on the ice fields to the east of Garibaldi volcano, causing increased melting and spring flooding. This in turn could threaten water supplies from Pitt Lake as well as fisheries on the Pitt River.

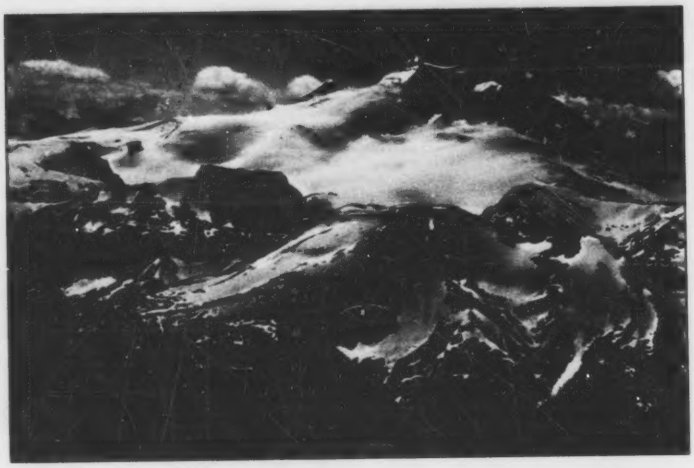


Figure A-4.

View looking south of four volcanoes with different morphologies. Garibaldi volcano, which encompasses three different mountain peaks (Mount Garibaldi, Atwell Peak, and Dalton Dome), is the large stratovolcano in the background. Table Mountain is a flat-topped, steep-sided volcano that erupted under glacial ice. Clinker Peak and Mount Price are the volcanoes in the foreground. Prominent levees are visible emanating from Mount Price. These levees demarcate the edges of a lava flow that erupted about 10 000 years ago. Photograph by C.J. Hickson. GSC 2000-138

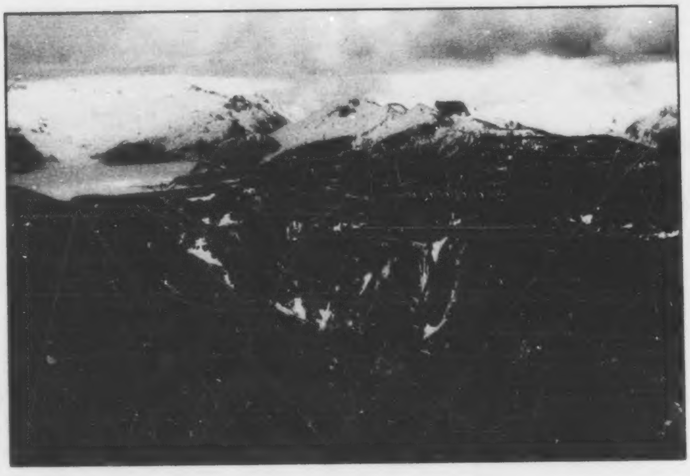


Figure A-5.

Lava flows emanating from Mount Price flowed towards the Cheakamus River valley (foreground). At the time of eruption, the valley was filled by glacial ice. The lava flow was stopped by the ice and ponded, eventually cooling. When the ice melted away, the ice-cooled lava-flow front formed a precipitous cliff; water ponded behind the lava dam, forming Garibaldi Lake (ice-covered lake to the left). The steep cliff failed in a series of landslides, the most recent of which occurred in 1855–1856. The location is referred to as 'The Barrier'. View to the east-southeast. Photograph by C.J. Hickson. GSC 2000-139

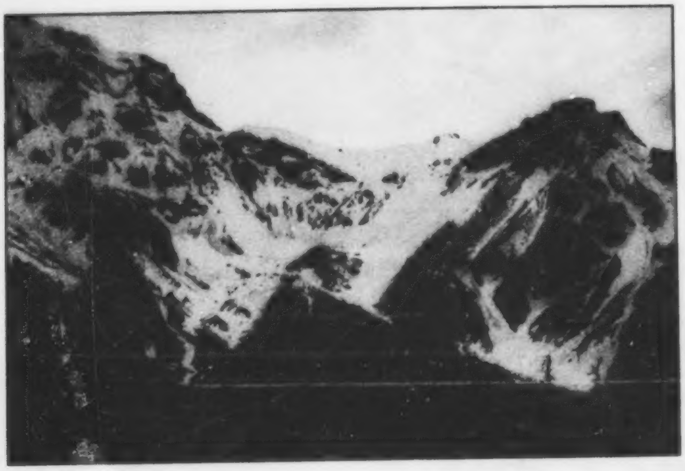


Figure A-6.

The eruption of Mount Meager volcano that occurred in 2350 BP emanated from the top of the dark triangular area in the centre of this photograph taken looking westward. It sent a column of ash at least 20 km high above the volcano. Ash was carried downwind as far as Alberta, and nearby regions were devastated by heavy ash fall, pyroclastic flows, collapsing lava flows, and catastrophic flooding. Photograph by C.J. Hickson. GSC 2000-140

Mount Meager (southwestern British Columbia)
Location: 48.79°N, 121.82°W
Volcanic features: Stratovolcano, lava flows, pyroclastic flow and fall deposits, flood deposits, hot springs
Oldest known eruption: 2.2 million years ago
Most recent known eruption: 2350 radiocarbon years ago
Summit elevation: 2645 m
Base elevation: 460 m
Estimated volume erupted: (?)<10 km³

Mount Meager, 150 km north of Vancouver, British Columbia, is the youngest of four overlapping stratovolcanoes resting on a 400 m high ridge of nonvolcanic, crystalline and metamorphic rock. The complex comprises at least eight vents that have produced basaltic to more evolved andesitic, dacitic, and rhyolitic magmas. Andesite lava flows (500 000 to 1 000 000 years old) are the most abundant rock type and their maximum total flow thickness is over 1 km. The most recent volcanic activity started 2350 radiocarbon years ago from a vent on the northeast side of the mountain and consisted of a massive, dacitic, Plinian eruption (Fig. A-6). The ash plume from this explosion travelled east across British Columbia into Alberta. The next stage of the eruption was marked by the production of hot, mobile, pyroclastic flows that travelled 7 km downstream from the vent (Fig. A-7, F-8). This early phase was followed by the extrusion of a lava flow (Fig. A-9) that repeatedly collapsed on the steep slopes, producing a unique, dense, welded breccia thick enough and extensive enough to dam the Lillooet River valley, resulting in the formation of a lake upstream along the Lillooet River. One dam failure on the upper Lilloet River produced a cataclysmic outburst flood that sent house-sized boulders downvalley for several kilometres and destructive floodwaters even farther. The final stages of the eruption produced a short lava flow. Two peripheral clusters of hot springs are found within the complex and suggest that magmatic heat is still present. Canada's only producing pumice mine is on the northeast side of Mount Meager.

***Figure A-7.** The initial vent-clearing Plinian eruption at Mount Meager deposited ash several metre in thickness close to the vent area. The black vertical objects seen to the right and above the person are carbonized tree trunks. The trees were first buried by falling ash (the layer extending from the person's feet to his waist), then were overcome by even thicker accumulations of pumice blocks from at least two pyroclastic flows. Above the pyroclastic-flow deposit is a welded breccia unit formed from the collapsing front of the lava flow that emanated from the vent area on the slopes far above. Photograph courtesy of P. Read*

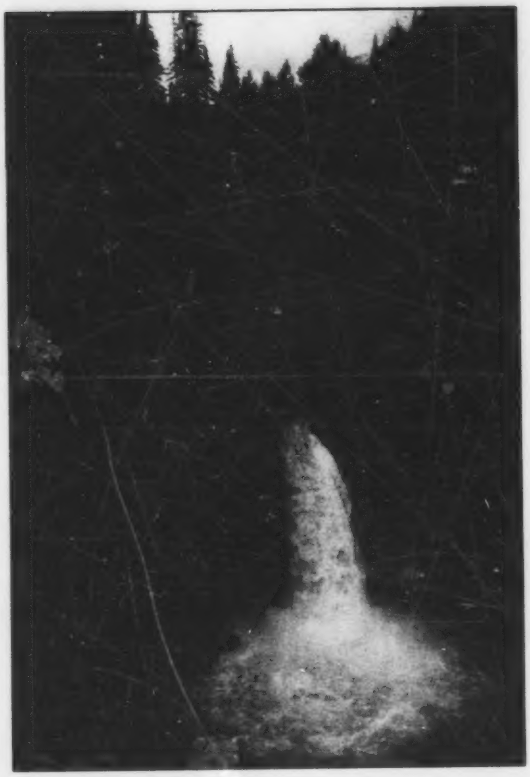


Figure A-8. This solid-looking rock cliff actually started as an unconsolidated volcanic breccia. Formed from the collapsing lava-flow front, the breccia collected downslope from the vent in the Lillooet River valley. The hot breccia accumulated so rapidly that it welded into a solid unit. This unit formed a rock dam behind which water from the Lillooet River ponded. Later breccia was not so well consolidated and the rising flood waters soon broke through the overlying material. The flood was massive enough that house-sized blocks of the underlying welded breccia were carried away as the flood eroded downward through the deposit. The Lillooet River has now cut a notch through the unit, forming Keyhole Falls. Photograph by C.J. Hickson. GSC 2000-141

The explosive nature of past eruptions at Mount Meager suggests that this volcano poses a significant long-distance threat to communities across southern British Columbia and Alberta. It also poses a significant local threat to the town of Pemberton, British Columbia, about 50 km downstream from the volcano. The failure of natural dams formed on the upper Lillooet River by landslides not associated with volcanic activity could produce catastophic floods. In addition, volcano-generated debris flows (lahars) could travel as far as Pemberton. Another explosive eruption at Mount Meager would also have a significant impact on local mining and logging operations, as well as significant negative impacts on the Lillooet River fishery (the river has salmon runs at least to Lillooet Lake, if not farther west).

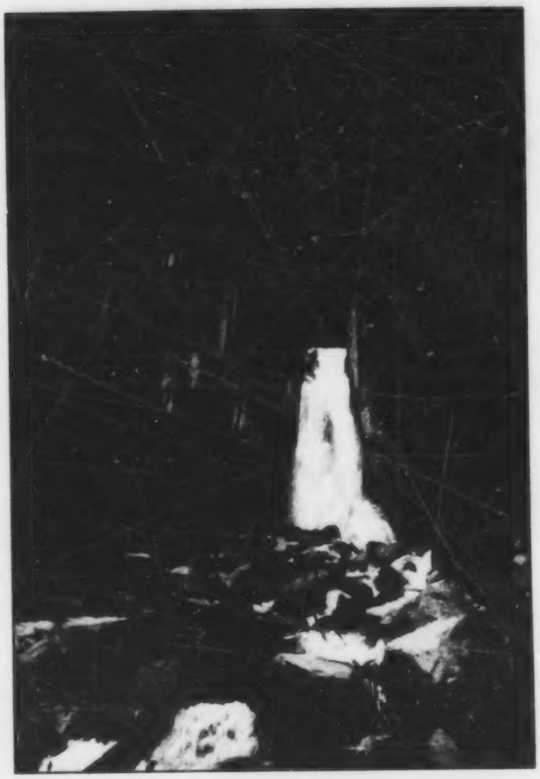


Figure A-9. A short, stubby lava flow was the last volcanic event to rock the Mount Meager area. The dacitic lava cooled into well formed columns over and through which Fall Creek has cut a path. Photograph courtesy of P. Read

Wells Gray volcanoes (east-central British Columbia)
Location: 48.79°N, 121.82°W
Volcanic features: Volcanic field of cinder cones and subglacial mounds
Oldest known eruption: 3 million years ago
Most recent known eruption: ~400 years ago
Summit elevation: 2100 m (highest volcanic features)
Base elevation: 600 m (lowest volcanic features)
Estimated volume erupted: 25 km³

The Wells Gray region of east-central British Columbia is a volcanic field made up of numerous, small, basaltic volcanoes. Individual volcanoes have been active for at least the last 3 million years during which time the region was covered by thick glacial ice at least twice, prior to the well known Fraser Glaciation (also known as the 'Wisconsin

Glaciation'). Volcanic eruptions underneath and through the thick blankets of glacial ice produced numerous unique glacial volcanoes and deposits, including one explosive, subaqueous volcano, five tuyas, at least one subglacial mound, and numerous, thick, valley-filling deposits of volcanic rocks. Between the periods of glaciation, the volcanoes continued to erupt and filled river valleys with many layers of basaltic lava flows. The valley-filling flows, now incised by rivers, have created a unique landscape of steep, narrow valleys and numerous waterfalls. A particularly beautiful waterfall is the 142 m (465 ft.) high, free-falling Helmcken Falls (Fig. A-10). Its beauty and that of the surrounding area prompted the provincial government to set the area aside as a park in 1939.

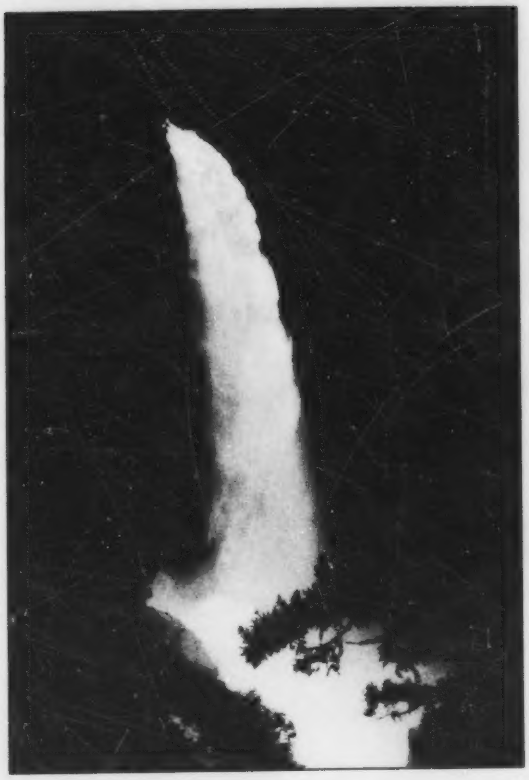


Figure A-10. Wells Gray Park owes its origin to the 142 m free-falling Helmcken Falls. The Helmcken Falls owe their origin to numerous lava flows that filled up the ancient channel of the Murtle River. At the end of the last ice age, massive floods from the melting ice carved a deep canyon into the underlying lava flows and formed Helmcken Falls. The face of the basaltic flows and falls remains vertical due to the nature of the basaltic flows. Basaltic lava shrinks as it cools and forms vertical columns. Rows of columns peal of the cliff face like slices of bread, preserving the vertical nature of the cliff. Photograph by C.J. Hickson. GSC 2000-142

A single road enters the park, but from it, a number of the park's volcanic features can be viewed. Short hikes lead to several spectacular volcanic features. The road terminates at the southern end of Clearwater Lake, where a short hike up the Kostal Lake trail leads to the Dragon's tongue lava flow (Fig. A-11). The flow, which is now punctuated by tree moulds visible from the trail, travelled 15 km from the vent. It forms a 3 m high dam across the southern end of Clearwater Lake, which maintains the existing lake level. This flow is just one of the examples of volcanic activity that have occurred in the region since the last ice age. This activity involved small fire-fountaining eruptions that produced cinder cones and lava flows. Other cones in the area include Kostal cone (at a few hundred years old, it is most likely the youngest in the region), Flourmill cone, and Spanish cones. Some lava flows at Wells Gray Park are unique in that they contain small, angular to rounded fragments of rock (called 'nodules') and crystals that come from the mantle below the Earth's crust. These green nodules are called 'peridotites' because they are composed predominantly of the mineral olivine (the gem form of which is known as 'peridot'). These lavas also contain large crystals of olivine, plagioclase, and pyroxene that crystallized deep within the Earth's crust and mantle. The lavas and the nodules they contain are similar to those erupted in the Volcano Mountain area of the Yukon Territory. The nodules help scientists determine what the mantle below the volcano is like.

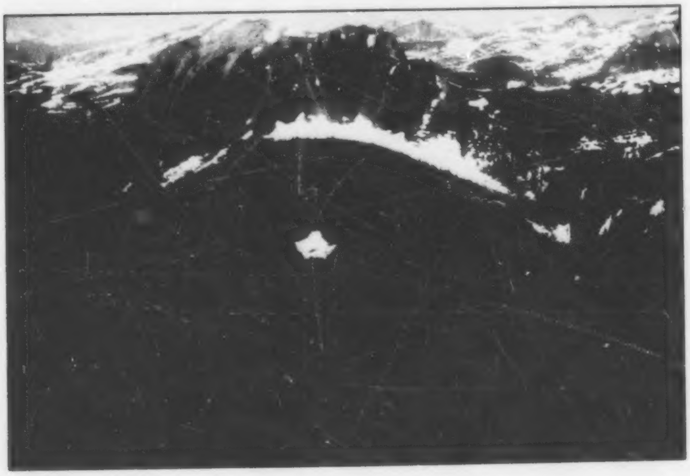


Figure A-11.

Dragon cone is the source of the 15 km long Dragon's tongue lava flow that dams Clearwater Lake. The cone, made up of cinders, blocks, and bombs, is perched on the side of a ridge of metamorphic rock. Photograph by C.J. Hickson. GSC 2000-143

A viewing platform on Green Mountain in the southern part of the park affords views of a number of the park's tuyas including McLeod Hill, Mosquito Mound, and Pyramid Mountain. McLeod Hill and Mosquito Mound are prominent, flat-topped, steep-sided volcanoes with a typical tuya shape (*see* Fig. 18, stage 3b). Pyramid Mountain (Fig. A-18), with its pyramidal shape, is a subglacial mound that never broke through the overlying ice cap (*see* Fig. 18, stage 2b).

White Horse Bluff, another distinctive subglacial volcano at Wells Gray Park, has a most unvolcano-like form (Fig. A-13). The bluff is an eroded remnant that marks the place of a violent eruption, thought to have involved the following sequence of events: 1) water, likely dammed by glacial ice, filled the Clearwater River valley; 2) the water flooded down the volcano's vent, producing large explosions of steam and broken lava fragments; 3) once the explosions had subsided, these fragments settled back into the water, building up a volcano composed almost entirely of fragmental volcanic glass (hyaloclastite).

Wells Gray Park has had a long history of relatively quiet fire-fountaining-style eruptions. More violent explosions are possible only in unique circumstances, such as an eruption into a lake. Any future eruption is most likely to affect only a limited area downslope from the volcano. Poisonous gases could be expelled and, as the flows are likely to travel long distances, it is conceivable that drainage systems could be rearranged and some river valleys might be dammed. Lava flows could start forest fires as most of the region is wooded. Salmon do not swim very far into the park, but future eruptions could have an impact on the significant recreational trout fishery in the park's lakes and river. Disruption of local air traffic and tourism is possible and could represent a crowd-control hazard in this remote yet relatively accessible region.

Nazko cone (central British Columbia)
Location: 52.90°N, 123.73°W
Volcanic features: Cinder cone, lava flows, and air-fall tephra
Oldest known eruption: 340 000 years ago
Most recent known eruption: 7200 radiocarbon years ago
Summit elevation: 1200 m
Base elevation: 1066 m
Estimated volume erupted: <0.1 km³

Figure A-12.

Pyramid Mountain, once mistaken for a cinder cone, is now known to have formed below several thousand metres of glacial ice. Hemmed in by the surrounding ice, the volcano erupted vigorously, creating layers of glassy, but vesicular, scoria intermixed with smoothed cobbles of granite and metamorphic rock. These rocks were picked up by a glacier many kilometres away and melted out of the ice because of heat from the volcano. Although it had a vigorous start, the eruption that formed Pyramid Mountain was not sufficiently sustained to form a larger edifice that could break through the surrounding ice and water to form a tuya. Photograph by C.J. Hickson. GSC 2000-144

Figure A-13.

White Horse Bluff is a deceptive volcano. Only a trained eye can pick out the yellow-weathering hyaloclastite and recognize that this is indeed a volcano. Formed by repeated subaqueous explosions, the volcano eventually built up above the water level. It was fed by dykes (dark areas on the cliff face) cutting through the hyaloclastite and ceased erupting soon after breeching the surface of the water. Photograph by C.J. Hickson. GSC 2000-145

Nazko cone is a small, tree-covered volcano in central British Columbia, approximately 75 km west of Quesnel. It is believed to be the easternmost expression of the Anahim hot spot that stretches across central British Columbia from Nazko to the Pacific Ocean. The volcano comprises lava flows, a composite cinder cone, and air-fall tephra. The sequence of eruption at Nazko started with the eruption of two distinct series of fluid lava flows, an older, grey basalt overlain by a younger, darker black basalt. The passive lava eruptions were followed by a period of explosive, pyroclastic eruptions of volcanic bombs and ash. This explosive activity built three overlapping cinder cones that were breached by the eruption of two lava flows near the ending stage of the explosive eruptions. Finally, explosive volcanic activity scattered tephra to the north and east of the cones. The tephra deposits are deepest near the cones (>3 m) and thin to less than a few centimetres only a few kilometres away, suggesting that the explosive eruptions were relatively small (Fig. A-14). However, scientists at the Geological Survey of Canada surmise that this last eruption at Nazko cone may have started forest fires because charcoal is found within the tephra layer.

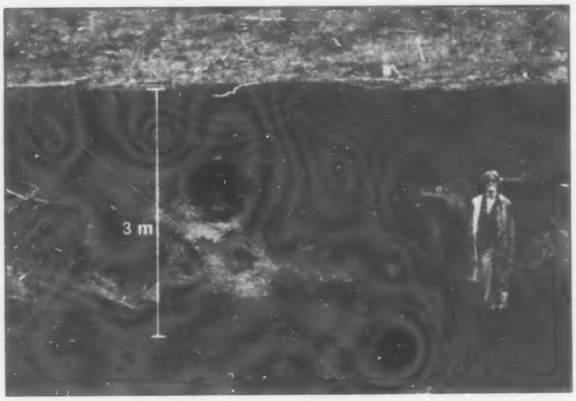


Figure A-14. The accumulations of tephra from Nazko cone are over several metres thick in some places. In this photo, the tephra is 2 m thick and unweathered and unconsolidated, making it a prime target for commercial exploitation. Photograph by C.J. Hickson. GSC 2000-146

The lowermost (hence oldest) lava flows at Nazko volcano have been dated at approximately 340 000 years using the K-Ar method. The youngest deposits of air-fall tephra were dated indirectly by analyzing burned wood within the tephra; the burned wood ranges from 10 000 to 7000 radiocarbon years old. The ages for the tephra can be confirmed geologically because the air-fall material is found on top of glacial till deposited during the Fraser Glaciation, which ended about 10 000 years ago. The cinder cones at Nazko are currently mined for their valuable resources of scoria, which is used for light-weight aggregate, landscaping and groundcover, and in agricultural and horticultural applications as a soil additive (Fig. A-15).

The most immediate hazard relating to future eruptions from Nazko cone is of local concern only and includes the possibility of forest fires and the disruption of local air traffic should an ash cloud be produced.

Tseax cone, Aiyansh, Nisga'a Memorial Lava Beds (central British Columbia)
Location: 55.12°N, 128.90°W
Volcanic features: Cinder cones and lava flows
Oldest known eruption: 625 radiocarbon years ago
Most eecent known eruption: 250 radiocarbon years ago
Summit elevation: 600 m
Base elevation: ~200 m
Estimated volume erupted: ~0.5 km³

The Nisga'a Memorial Lava Beds Provincial Park is approximately 60 km north of Terrace, British Columbia, and contains cinder cones and accompanying lava flows of the Aiyansh– Ksi Sii Aks volcanos. The cinder cones are situated in a valley above and east of the Ksi Sii Aks, about 20 km south of the river's junction with the larger Nass River. The most recent cone, 290 m in diameter at the base, rests on the remains of an earlier and somewhat larger, dissected, 460 m

Figure A-15.

Nazko cone is easily reached by road by travelling west from the main highway running through the town of Quesnel, British Columbia. The fire-fountaining eruptions that built the cone produced thick accumulations of scoriaceous tephra, which covered the surrounding landscape. Here the scoria is being stripped to be used in a number of commercial applications. Photograph by C.J. Hickson. GSC 2000-147

diameter cone. The cones formed in the narrow confines of a tributary of the Ksi Sii Aks and comprise basaltic bombs and cinder. The most recent lava flowed 22.5 km from the vent to the Nass River where, according to legend of the Nisga'a people, it blocked the flow of the Nass River (Fig. A-16). Along the south edge of the Nass River, the exposed lava contains structures indicative of lava that has flowed over wet ground. The provincial park is a must for volcano enthusiasts because it is easily accessible and has a wide range of lava-flow features, including pahoehoe and aa lava flows.

The eruption of Tseax cone is the only eruption in Canada for which legends of First Nations people have been verified. The Nisga'a (Tsimshian people) tell of a prolonged period of disruption by the volcano, including the destruction of their village on the Nass River and the death of some people (reportedly 2000) from "poisonous smoke" (Barbeau, 1935, p. 223). The vent was active at least twice (220 and 650 years ago) and remnants of other, older lava flows exist in the area. Nisga'a Memorial Lava Beds Park became a reality in 1993 when the province of British Columbia set aside the area to honour the dead and to preserve the unique region.

If Tseax cone were to erupt again, modern monitoring techniques and an understanding of volcanic hazards would help prevent a repeat of the 'poisonous smoke' disaster. The poison smoke was most likely odourless, heavier than air, CO_2 gas. The CO_2 replaces air in depressions and poorly ventilated structures. People sleeping in these areas would have had no knowledge or warning of the danger. Because of our knowledge of this previous disaster, modern monitoring would include studies of the gases emitted by the volcano and a warning to people living downslope from the volcano. Future eruptions could start forest fires and potentially dam local rivers (as happened to the Ksi Sii Aks and Nass River) if the volume of lava were significant enough. If lava were to again reach the Nass River, it could have disastrous short-term consequences for the important salmon fisheries on the Nass River system.

Lava Fork (northwestern British Columbia–southeastern Alaska)
Location: 56.48°N, 130.77°W
Volcanic features: Scoria cone, air-fall deposits, and lava flows
Oldest known eruption: >350 years ago
Most recent known eruption: ~150 years ago
Summit elevation: 1200 m (highest volcanic feature)
Base elevation: 100 m (lowest volcanic feature)
Estimated volume erupted: 2.2 km³

Lava Fork volcano is located immediately north of the British Columbia–Alaska border in northwestern British Columbia, approximately 60 km northwest of the town of Stewart. The volcano is one of 10 found along or near the Iskut River. The deposits at Lava Fork comprise mainly basaltic lava flows (Fig. A-17), which erupted from a vent high up on the side of a glaciated, U-shaped valley and flowed south 5 km where they crossed the border into Alaska and dammed the Blue River, forming several small lakes. In total, the lava flows are approximately 22 km long (Fig. A-18). The vent area has small deposits of sulphur precipitated from volcanic gases and volcanic bombs up to 0.5 m long. The surface of the lava flows still have well preserved flow features such as pressure ridges and lava channels (Fig. A-18), as well as pits formed when the overlying solidified lava collapsed into underlying lava tubes. In many areas, large trees were engulfed by the lava flows and are now found embedded in the top of the flows (Fig. A-19).

At least two different episodes of eruption produced the basaltic lava flows at the Lava Fork vent. On the basis of tree-ring-core dating and ^{14}C dating, the youngest of these lava flows is probably only 150 years old. This makes Lava Fork the youngest known volcano in Canada. Because the Iskut region is so remote and the style of eruption involves basically passive, fluid lava flows, future eruptions from Lava Fork pose little threat. Damming of local water courses may disrupt fish habitat and spawning grounds, but ash clouds are unlikely to be high enough to disrupt international

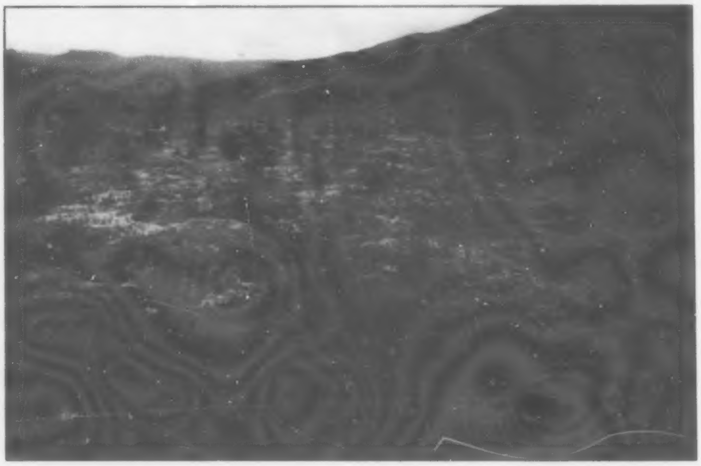


Figure A-16.

Lava flows from the Tseax cone ponded near the Nass River, forming a vast, flat plain underlain by the lava flows. The fluid lava, travelling over 22 km from the cone, filled in any preexisting topographic depressions, then spread out and solidified. The flows most likely dammed the Nass River for a short period of time. Photograph by C.J. Hickson. GSC 2000-148 jet traffic. They could potentially threaten lower flying aircraft along the northern coastal corridor between Vancouver and Alaska.

Iskut River (northwestern British Columbia)
Location: 48.79°N, 121.82°W
Volcanic features: Cinder cones, lava flows
Oldest known eruption: ~70 000 years ago
Most recent known eruption: <2555 radiocarbon years ago
Summit elevation: 610 m
Base elevation: 150 m
Estimated volume erupted: 8.6 km³ (maximum)

Cinder cones along the east and west sides of the Iskut River produced at least 12 basaltic lava flows in the last 70 000 years. The lava flows commonly have clinkery surfaces and locally preserved lava-flow features. Several of the Iskut River lava flows were large enough to block the river, which has subsequently cut down through the lava flows to form the Iskut River canyon, an extremely rugged, narrow gorge that has blocked the migration of salmon into the upper reaches of the Iskut River. The youngest Iskut River lava flows have not been dated successfully, but are younger than a flow dated at 2555 radiocarbon years before present. Several other young lava flows (<100 000 years old), which have not been dated directly, occupy tributary valleys of the Iskut River, including Snippaker Creek, Tom McKay Creek, King Creek, and Second Canyon. In the mid-1980s, the Iskut River canyon was studied as a possible site for a hydroelectric project. However, the project was abandoned in part because of the uncertain volcanic future in the area. Hazards associated with future eruptions in this unpopulated region are minor.

Hoodoo Mountain (northwestern British Columbia)
Location: 48.79°N, 121.82°W
Volcanic features: Stratovolcano, subglacial lava flows, pyroclastic deposits, aa lava flows, ice- dammed lava cliffs
Oldest known eruption: >100 000 years ago
Most recent known eruption: ~7000 years ago
Summit elevation: 1850 m
Base elevation: 100 m
Estimated volume erupted: 17 km³

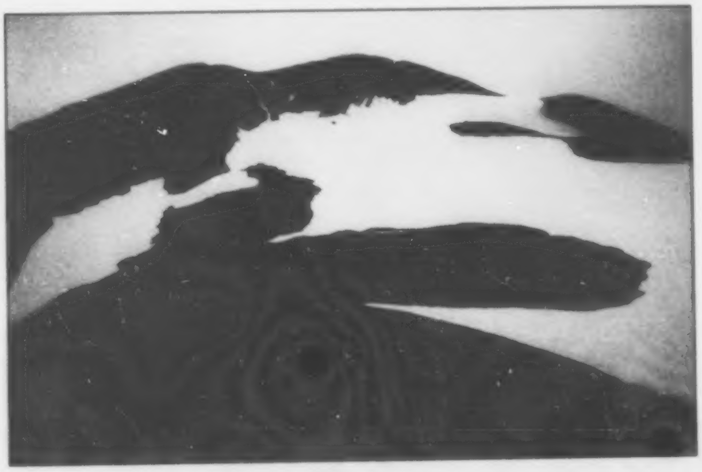


Figure A-17.

A helicopter sits upon thick, black scoria deposited from the eruption of Lava Fork volcano. The surrounding ridges are also covered with tephra and scoria deposits. The cone itself is poorly formed and, even though it is very young, has been eroded by the perpetual ice found at this elevation and latitude. Photograph courtesy of J.K. Russell.

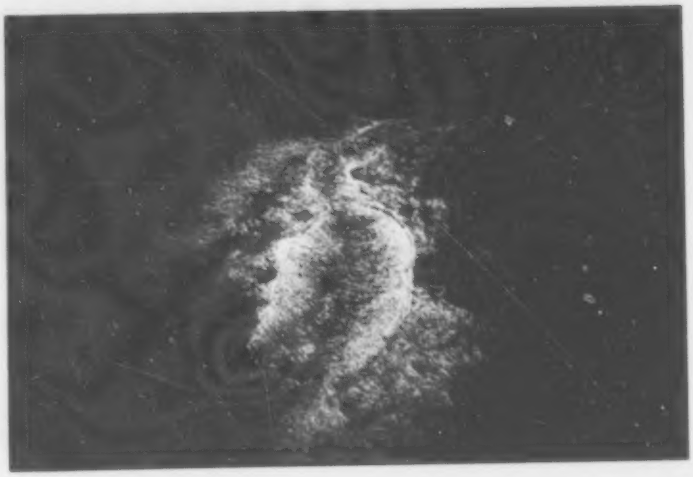


Figure A-18.

Well defined lava levees and flow features can be seen on the surface of lava flows emanating from Lava Fork volcano. Although the vent area is relatively nondescript, significant quantities of lava poured down the steep slopes into the valley below, travelling over 15 km. Photograph by C.J. Hickson. GSC 2000-149

Hoodoo Mountain is a flat-topped volcano on the north edge of the Iskut River, about 100 km northwest of the town of Stewart, in northwestern British Columbia (Fig. A-20). The volcano has an ice cap 3 km in diameter and, throughout most its history, has been influenced by glacial ice. Most of the volcanic deposits at Hoodoo Mountain are lava flows (Fig. A-21); however, some pyroclastic rocks are also found, indicating at least one period of explosive activity. Most of the volcano was formed beneath glacial ice, and it has been overlain by glaciers or an ice cap for most of its eruptive history. All recent flows exposed on its flanks appear to have originated from beneath the existinf summit-covering ice (Fig. A-22). The edifice consists of a volcanic pile of fine-grained lavas interbedded with glass-rich breccia (hyaloclastite), both welded and nonwelded pyroclastic breccias, and highly crystal-rich (porphyritic) lava flows. Several of the youngest porphyritic lava flows have not been heavily glaciated.

Hoodoo Mountain volcano is in an area currently (1997) mined for gold and wollastonite, and several large mining camps are within 15 km of its south flank. Should there be renewed activity in the form of lava flows, damming of the Iskut River could be a significant hazard to current mining operations along the river. Mining infrastructure is located within the floodplain of the river and could be damaged by rising waters. Similarly, eruptions through the summit ice cap or onto the surrounding glaciers could also cause significant flooding of the Iskut River and the lower portion of the larger Stikine River. Although not as disruptive to mining operations, flooding of the Iskut and Stikine rivers could seriously disrupt the salmon fishery on the Stikine River. Air-fall tephra from an explosive eruption could disrupt local mining operations and any airborne ash would disrupt air traffic to and from the mining camps. High ash columns would disrupt air traffic between Canada, Alaska, and Asia.

Mount Edziza (northwestern British Columbia)
Location: 57.72°N, 130.63°W
Volcanic features: Stratovolcanoes, shield volcanoes, cinder cones, calderas, aa and pahoehoe lava flows, subglacial eruption deposits, air-fall and pyroclastic flow deposits
Oldest known eruption: >7.5 million years ago
Most recent known eruption: 1350 radiocarbon years ago

Figure A-19.

Near the distal end of the Lava Fork flows, trees were engulfed by the laval flows, but preserved as lava quenched against massive trunks or groups of trees. The bases of the trees then burned, and the upper trunks and branches fell to the still-cooling lava surface below. These hulks can be seen littering the lower reaches of the flow in curious crisscross patterns. View looking south, in the flow direction; the large tree is lying parallel to the flow direction. Photograph by C.J. Hickson. GSC 2000-150

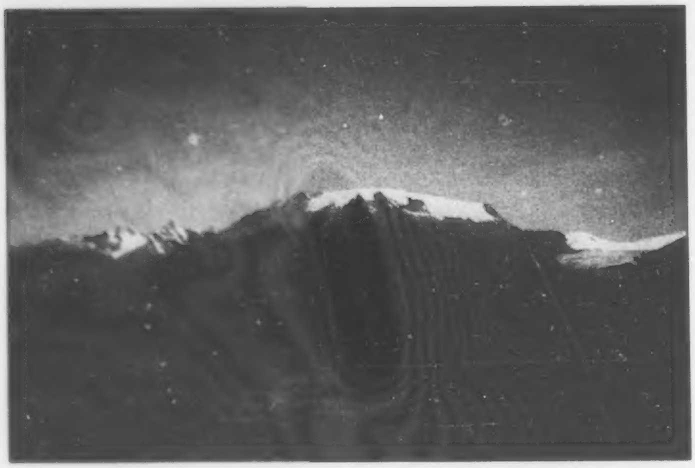


Figure A-20.

Hoodoo Mountain, viewed looking northwest, rises above the surrounding valley of the Iskut River and flanking glaciers on this rare clear day in the British Columbia Coast Mountains. The flattened top of the volcano attests to the volcano's constant struggle with surrounding and overlying ice during its 100 000 year eruptive history. Photograph by B.R. Edwards. GSC 2000-151

Summit elevation: 3285 m
Base elevation: 1650 m
Estimated volume erupted: 665 km³

Mount Edziza, the second largest volcanic complex in Canada, is in northwestern British Columbia. Its glacier covered peaks are visible from the Stewart–Cassiar highway, 40 km to the west. This area of long-lived volcanic activity was studied in detail for many years by Dr. Jack Souther of the Geological Survey of Canada. Souther's bulletin (1992) on Mount Edziza and the surrounding volcanoes highlighted the importance and size of the region, one of the largest areas of young volcanic activity in Canada. The complex comprises a large plateau of predominantly basaltic lava flows; four large stratovolcanoes and a shield volcano are built on top of the plateau. The plateau is about 65 km long and a maximum of 20 km wide. The higher peaks of the complex were built upon a base of fluid basalt lava flows and are now glacier covered (Fig. A-23). Throughout the length of the volcanic complex, numerous eruptions have occurred within the last 10 000 years (or since the end of the last ice age). The most recent activity formed two large lava fields and several smaller, isolated cinder cones surrounded by lava flows (Fig. A-24, F-25). Desolation lava field, covering 150 km² on the northern end of the volcanic plateau, is the largest area of young lava flows. The longest lava flow from the Desolation field, and the youngest in the volcanic complex, is 12 km long. Eve cone (Fig. 15), in the middle of the Desolation lava field, is one of the most symmetrical and best preserved cinder cones in Canada. Other young lava flows and cinder cones at Mount Edziza include Kana cone, Cinder cliff, Icefall cone, Ridge cone, Walkout Creek, the Mess Lake and Snowshoe lava fields, and two isolated cones at the southern end of the volcanic plateau. Many of the young cinder cones and lava flows at Mount Edziza are covered by light yellow pumice, referred to as the 'Sheep Track Member' by Souther (1992). The volcano that produced the pumice has not been discovered, but probably has been covered by glacier ice.

Active or recently active warm springs are found in several areas along the western flank of the volcanic plateau, including Elwyn springs (36°C), Taweh springs (46°C), and inactive springs near Mess Lake. All three hydrothermal

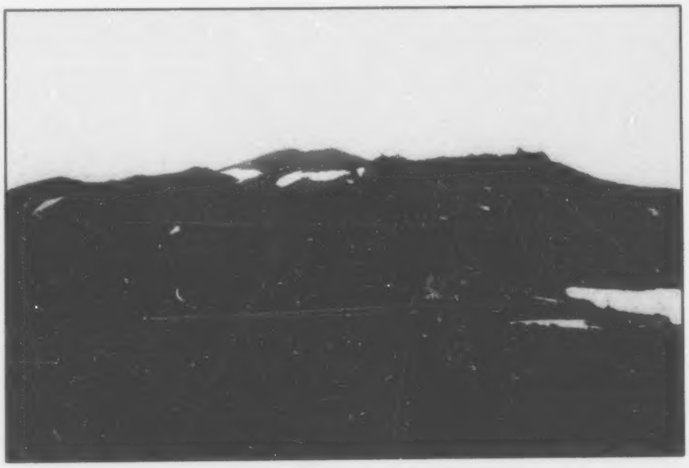


Figure A-21.

The youngest lava flows from Hoodoo Mountain emanate from beneath the overlying summit ice cap. The flows are very thin and have aa surfaces. Lava levees and other flow features are well preserved in may areas. Photograph by B.R. Edwards. GSC 2000-152

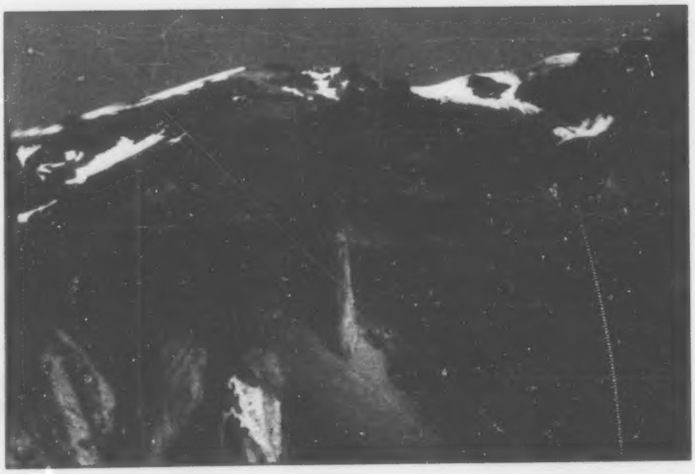


Figure A-22.

Thick lava flows that ponded against glacial ice form the flanks of Hoodoo Mountain. At least eight flows are exposed in this 500 m high cliff face. The thickest is 75 m. The prominent spire in the centre of the photograph is a more erosion-resistant dyke cutting through the flows. Photograph by B.R. Edwards. GSC 2000-153 areas are near the youngest lava fields on the plateau and are probably associated with the most recent volcanic activity at Mount Edziza. The vast plateau has also been an important cultural resource. The Tahltan people, who now live largely near Telegraph Creek, British Columbia, used volcanic glass (obsidian) from Mount Edziza to make tools and for trading material. More recently, most of the plateau has been made into a provincial park to preserve the volcanic and culture treasures unique to the northern British Columbia area.

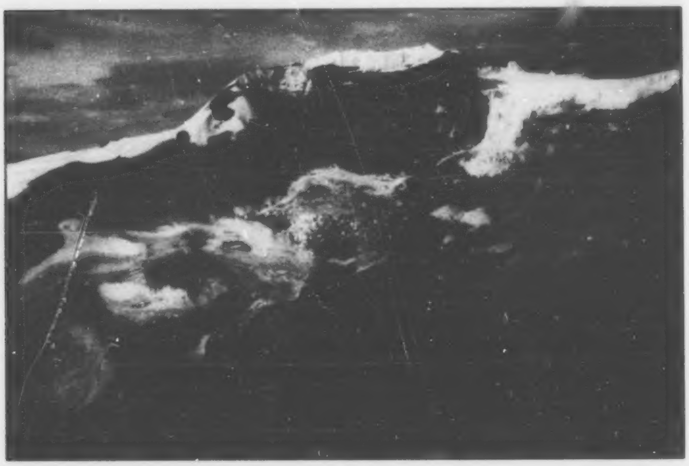


Figure A-23.

Mount Edziza is the remnant of a large shield volcano, now heavily eroded by overriding glaciers. The peak is 2590 m high. Photograph by C.J. Hickson. GSC 2000-154

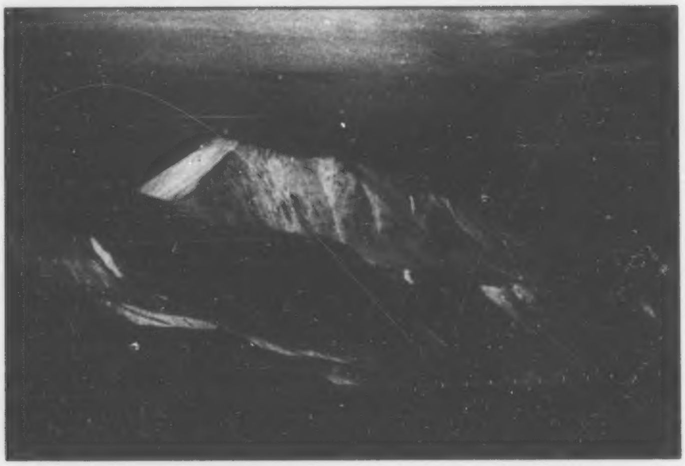


Figure A-24.

Pyramid Mountain is a relatively young dome on the northwest flank of Mount Edziza. The dome represents just one of the several different styles of volcanism found in the Mount Edziza volcanic complex. Photograph by C.J. Hickson. GSC 2000-155

Figure A-25.

Williams cone is one of numerous postglacial cinder cones dotting the landscape around Mount Edziza. The most recent of these has been dated at 1350 radiocarbon years ago. Photograph by C.J. Hickson. GSC 2000-156

The Sheep Track Bench pumice deposit highlights one of the important volcanic hazards associated with the Mount Edziza volcanic complex — the possibility a large, explosive volcanic eruption. Mount Edziza has a history of producing not only fluid and relatively passive basaltic lava flows, but also more silica-rich trachytic and rhyolitic lava flows and explosive eruptions. The silica-rich compositions are similar to those associated with the most violent eruptions on Earth, and scientists believe that the ice cap on top of Mount Edziza might be covering at least one volcanic caldera or crater. An explosive eruption could produce an ash cloud that would affect parts of northwestern Canada. As well, in the past, lava flows from Mount Edziza and closely related volcanoes to the north have dammed the Stikine River, which now has a major salmon fishery. Forest fires started by volcanic eruptions are also a possibility in this part of the world.

Mount Churchill (Wrangell Mountains, Alaska)
Location: 61.25°N, 141.70°W
Volcanic features: Stratovolcano, caldera
Oldest known eruption: >1850 radiocarbon years ago
Most recent known eruption: ~1250 radiocarbon years ago
Summit elevation: 4766 m
Base elevation: ~900 m
Estimated volume erupted: >25–50 km³

Mount Churchill is a glacier-covered volcano in the Wrangell Mountains of Alaska and the Yukon Territory. Extreme elevations, remoteness, and a prolific ice cover have made investigations of this area difficult. The mountain is thought to be constructed primarily of young volcanic rocks, although most of these rocks are buried beneath a glacial cap that covers much of the mountain (Fig. A-26). The summit crater rim is composed chiefly of ash, coarse pumice, and rock fragments. The rock debris is interpreted to have been ripped

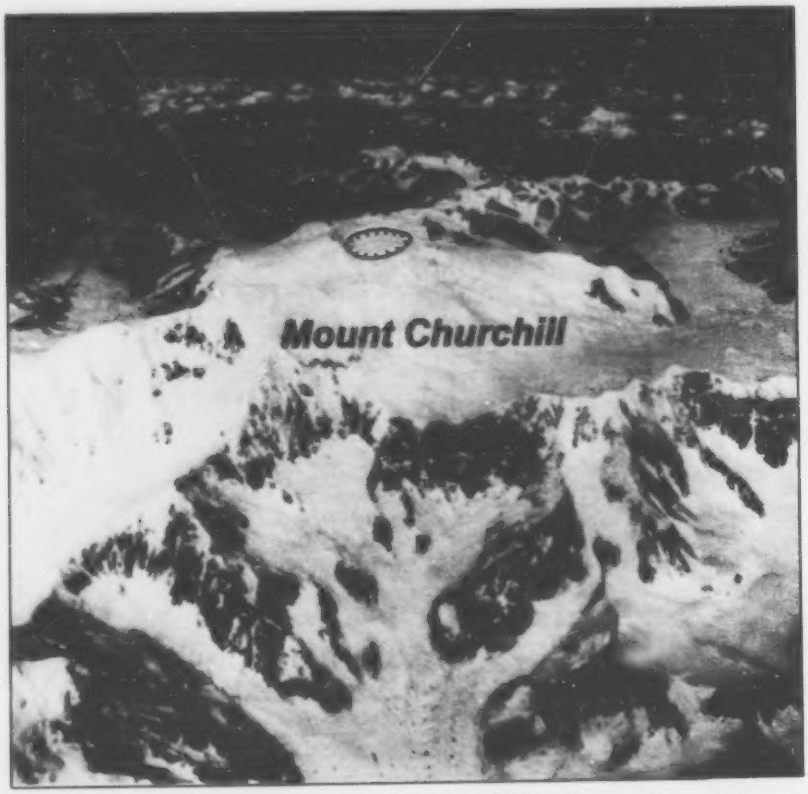


Figure A-26. Aerial view of Mount Churchill looking northward. A hypothetical caldera shown in hatched pattern is the currently proposed site of the eruption that produced the White River Ash. Photograph courtesy of R. McGimsey, United States Geological Survey from the volcano's conduit walls during Mount Churchill's explosive eruptions. Mount Churchill is the most likely source of the White River Ash, a thick layer of pumice from the most voluminous pyroclastic eruptions in North America in the last 2000 years (Fig. A-27). This ash deposit contains an estimated 25 to 50 km³ of tephra and covers over 340 000 km² in northwestern Canada and adjoining eastern Alaska. The pumice deposits form two lobes. Radiocarbon dating indicates that the northern lobe was deposited about 1890 radiocarbon years ago and the larger eastern lobe, about 1250 radiocarbon years ago. Field and laboratory studies strongly suggest that both lobes emanated from a common source vent.

The current distribution of White River Ash shows that another eruption from Mount Churchill, similar to past eruptions, could have widespread effects throughout the Yukon Territory and the southern half of the Northwest Territories. Remembering that the amount of ash preserved is probably much less than the true amount of ash erupted, we conclude that another eruption of similar magnitude would have a significant economic impact on Canada. Hazards would include ash damage to mechanized equipment, impact on forests from heavy ash accumulations, and heavy ash sedimentation in streams. Respiratory problems in people would increase significantly and the impact on animal populations could be catastrophic. All air traffic in northwestern Canada and eastern Alaska would be affected. Unconfirmed aboriginal legends from the area suggest that the eruption that occurred 1250 radiocarbon years ago disrupted food supplies and forced aboriginal people to migrate southward.

Volcano Mountain (central Yukon Territory)
Location: 62.93°N, 137.38°W
Volcanic features: Cinder cones and lava flows
Oldest known eruption: ~1 million years ago
Most recent known eruption: ~7000 radiocarbon years ago
Summit elevation: 1238 m
Base elevation: 913 m
Estimated volume erupted: ~0.05 km³

Volcano Mountain is located immediately north of the junction of the Pelly and Yukon rivers, half way between Whitehorse and Dawson City, central Yukon Territory. The volcano comprises a cinder cone and a series of lava flows

Figure A-27.

Mount Churchill is the most likely source of the White River Ash, a thick layer of white pumice visible in the upper part of this photograph. Photograph by L. Jackson. GSC 2000-181

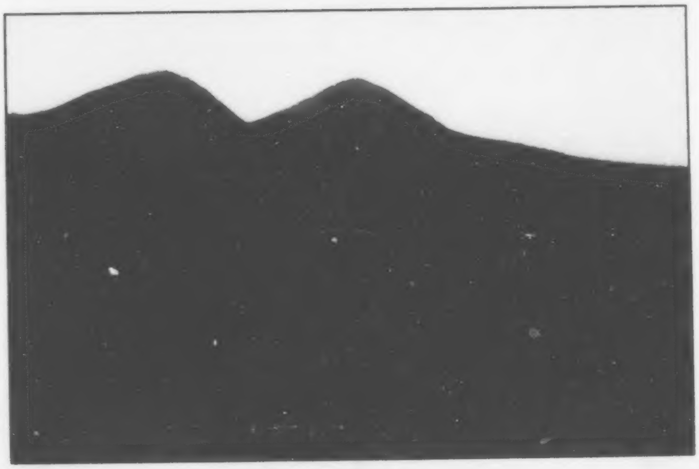


Figure A-28.

Volcano Mountain is a small, asymmetrical cinder cone approximately 300 m high. Lava flows issued through breaks in the cone wall to the northeast and to the southwest. Photograph by L. Jackson. GSC 2000-157

Figure A-29.

The surface of this basaltic lava flow from Volcano Mountain broke into chunks as the lava flowed. These chunks have been tossed about and upended creating a clinkery surface typical of some basaltic lava flows. Photograph by L. Jackson. GSC 2000-158

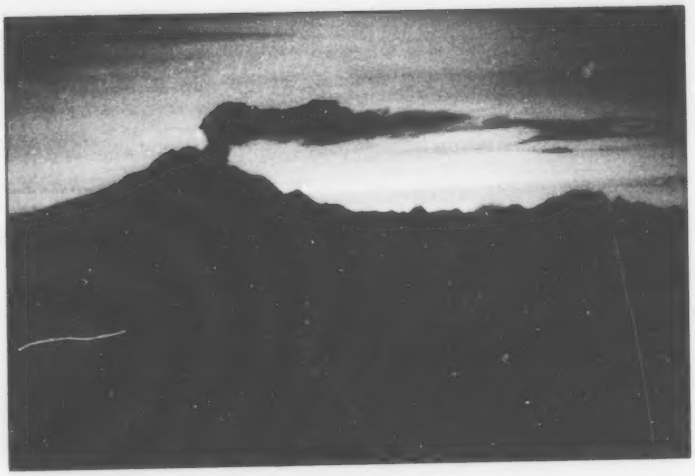


Figure A-30.

Mount Redoubt, a large stratovolcano along the north side of Cook Inlet, Alaska. Photograph courtesy of P. Hickson.

that issued through breaks in the cone wall both to the northeast and the southwest (Fig. A-28, F-29). The northeastern lava flow travelled about 5.5 km and the southwestern flows, only about 3 km. The lavas at Volcano Mountain represent a unique type of lava, called 'olivine nephelinite', that is very uncommon in the geological record. These lavas are generally thought to have come from much deeper in the Earth than the basalt magmas common to most of the other young volcanoes in northwestern British Columbia and the Yukon Territory. They are also unusual because they contain small, angular to rounded fragments (nodules) of rock that come from the mantle, below the Earth's crust. These green nodules are called 'peridotites' because they are composed predominantly of the mineral olivine or, in gem form, 'peridot'. These lavas and the nodules they contain are similar to those erupted in the Wells Gray area of British Columbia.

The fluid nature of the lavas and the lack of significant fragmental material erupted from Volcano Mountain suggest that any future eruptions would probably be dominated by lava flows, with little to no explosive activity. Thus, the main hazards posed by any renewed activity include forest fires started by the hot lavas. However, evidence from older volcanic deposits immediately south of Volcano Mountain, in the Fort Selkirk area, indicates that lava flows may have once partly blocked or at least altered the course of the Yukon and Pelly rivers. Renewed activity in this area could disrupt the course of one or both of these major rivers and have a serious impacts on people living or working downstream.

Other selected American volcanoes

Mount Redoubt (Fig. A-30) is a large stratovolcano along the north side of Cook Inlet, Alaska. During its most recent phase of activity, which started in late 1989, an explosive eruption formed an ash plume high enough to carry ash into the flight paths followed by many jet aircraft travelling between Asia and northwestern North America. The most significant event for Canadian volcanology was a near disaster when a 747 jet encountered the ash plume. The jet landed safely at Anchorage airport, but sustained significant damage — close to its total value. More importantly, the lives of several hundred people on board the flight were put at risk. As a result of the mishap, the aviation industry around the world called upon governments to implement better procedures to deal with volcanic ash. Canada was no exception. A multi-agency working group was convened in 1990 and the Interagency Volcanic Event Notification Plan was born shortly thereafter. The plan established notification procedures to alert all necessary government agencies during a volcanic eruption crisis that threatens Canadians.

Volcanologists forecast that **Mount St. Helens** in Washington State would erupt before the end of the twentieth century (Fig. A-31). That forecast proved correct in March 1980 when seismicity underneath the mountain began to increase, followed by steam explosions from the summit area. This precursor activity culminated at 08:32 h on 18 May 1980, with an explosive Plinian eruption. Explosive activity, interspersed with dome growth, continued at Mount St. Helens over the next six years.

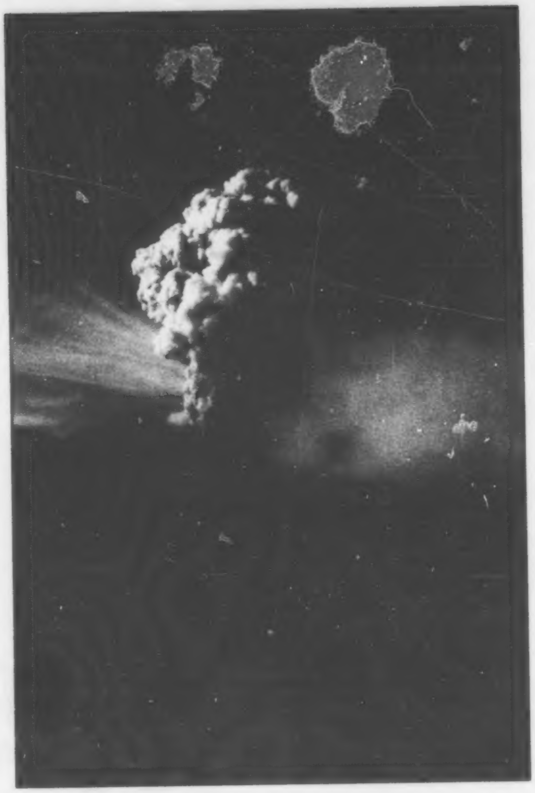


***Figure A-31.** Photograph of a small explosive eruption of Mount St. Helens that occurred in October 1980. Photograph courtesy of P. Hickson.*

Geomagnetic hazards

D.H. Boteler[1]

Boteler, D.H., 2001: Geomagnetic hazards; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 183–206.*

Abstract: This report provides a review of the effects of geomagnetic disturbances on technological systems on the ground. Information is provided on the causes and characteristics of different types of magnetic disturbances, and the factors affecting their occurrence are used to build up a geomagnetic climatology. This information is then used to make a hazard assessment for affected systems in Canada. Geomagnetic effects on power systems, pipelines, and telephone cables are examined in detail. A catalogue is provided of the 150 years of geomagnetic effects that have been observed on technological systems on the ground, and reports are given on using the celestial power.

Résumé : Le présent article passe en revue les incidences des perturbations du champ magnétique terrestre sur les systèmes techniques situés au sol et présente des données sur les causes et les caractéristiques des différents types de perturbations. Les facteurs qui engendrent la venue de ces perturbations sont utilisés pour l'élaboration d'une climatologie géomagnétique. Cette information est utilisée par la suite pour faire l'évaluation des risques que peuvent subir un certain nombre de systèmes au Canada. Les incidences des perturbations magnétiques sur les systèmes électriques, les pipelines et les câbles téléphoniques sont examinées en profondeur. Un catalogue des incidences des anomalies magnétiques observées sur les systèmes techniques au sol pendant 150 ans est inclus et des rapports sont présentés sur l'utilisation de l'énergie associée aux aurores.

---

[1] Geomagnetic Laboratory, Geological Survey of Canada, 7 Observatory Crescent, Ottawa, Ontario K1A 0Y3

INTRODUCTION

Since the introduction of electrotechnology in the middle of the nineteenth century, electrically conducting networks have been constructed over large parts of the globe. The telegraph system was the first, followed by telephone cables, both on land and along the seafloor. In the twentieth century, more and more pipelines have been built to bring oil and gas from distant locations. Electric power networks also use long transmission lines to carry electrical power from the generating stations to the cities where the power is used.

All these technological systems are enveloped by the Earth's magnetic field and be affected by electric fields induced by changes in the magnetic field. When solar disturbances cause magnetic storms, these magnetic-field variations become extreme and the induced electric fields can have a significant effect on the operation of these systems.

Geomagnetic effects on electrical systems have been observed for over 150 years. The early telegraph system was the first to be affected and many reports were written about the 'celestial voltages' making it impossible to send messages. In more recent times, 'geomagnetically induced currents', or 'telluric currents' as they are also called, have been observed in high-voltage power systems and in steel pipelines. In pipelines, the currents are a possible source of corrosion and also interfere with pipeline potential surveys that are part of corrosion prevention. In power systems, the currents have caused lines to trip out of service, transformers to overheat, and power blackouts for entire regions.

This contribution reviews the effects of geomagnetic disturbances on technological systems on the ground. It provides information on the causes and characteristics of different types of magnetic disturbances and uses the factors affecting their occurrence to build a geomagnetic climatology. This information is then used to make a hazard assessment for affected systems in Canada. Geomagnetic effects on power systems, pipelines, and telephone cables are examined in detail. Appendix A is a catalogue of the 150 years of geomagnetic effects that have been observed on technological systems on the ground, and Appendix B presents reports on using the celestial power.

MAGNETIC DISTURBANCES

From the earliest accounts, people have recognized the simultaneous occurrence of telluric currents on telegraph systems and fluctuations of compass needles and sightings of the aurora. Some also noted an association between telegraph disturbances and the number of sunspots crossing the face of the sun. The nature of the links between these phenomena has long been debated. Just as the requirements of navigators spurred research into the Earth's main magnetic field, disturbances on the telegraph helped stimulate investigations of the magnetic field disturbances.

The Earth's main magnetic field and magnetic disturbances are produced by radically different phenomena. The main field is of internal origin and is generated by electric currents flowing in the Earth's molten outer core. Heat convection causes the molten iron within the core to cirulate. The movement of the conducting iron through the Earth's magnetic field is presumed to generate an electric current that flows in such a way as to reinforce the magnetic field.

Magnetic disturbances are produced by electric currents flowing in the ionosphere and magnetosphere, which extend from 100 km above the Earth's surface to many thousands of kilometres out into space. Every electric current produces a magnetic field, and the magnetic field of currents are observed at the Earth's surface along with the Earth's own magnetic field. Thus a geomagnetic disturbance is not a disturbance in the Earth's own magnetic field, but rather the superposition of the 'external' magnetic field produced by ionospheric and magnetospheric currents.

External electric currents fluctuate, so the magnetic fields they produce also fluctuate. Any electrical conductor in a varying magnetic field will be affected by an induced electric field. Thus, man-made conductors, such as pipelines, telephone cables, and power transmission lines, at the Earth's surface will be affected by this electric field. However, the Earth is also an electrical conductor and electric currents are induced within it. This complicates things, because the induced currents in the Earth also create a magnetic field. Thus, the total magnetic disturbance affecting electrical conductors at the Earth's surface comprises an external part due to currents produced in the ionosphere and magnetosphere and an 'internal' part due to currents induced in the Earth.

The phenomema that give rise to geomagnetic disturbances on Earth originate on the sun. The sun radiates charged particles (mainly electrons and protons) into space, called the 'solar wind', that travel outward into space. When the solar wind reaches Earth, it is deflected by Earth's magnetic field, which is itself compressed on the sunward side and drawn out into a comet-like tail on the night side. The region of space that is shielded from the solar wind by the Earth's magnetic field is known as the 'magnetosphere' (Fig. 1). The compression of the magnetic field on the sunward side of Earth is produced by electric currents on the 'magnetopause', the outer edge of the magnetosphere, which cancel Earth's magnetic field outside the magnetosphere and add to it inside.

Eruptions on the sun also produce high-speed clouds of particles within the solar wind. When these high-speed particles reach Earth, they can cause a magnetic storm (Fig. 2). Some storms start with a 'sudden commencement', a sharp jump in the magnetic field due to the shock wave produced by the impact of the particle cloud on the magnetosphere. During the initial phase of the magnetic storm, the high-speed solar wind compresses the magnetosphere, causing an increase in the magnetic field at the Earth's surface.

In the main phase of the magnetic storm, the interaction between the solar wind and the magnetosphere leads to charged particles becoming trapped within regions of the magnetosphere, called the 'Van Allen belts', 25 000 to 40 000 km above Earth. These particles oscillate around the magnetic-field lines and in doing so, drift around Earth, electrons moving eastward and protons moving westward. This produces a ring current around Earth that causes a depression of the horizontal magnetic field seen all around the world. During the recovery phase, which lasts one day or more, the ring current subsides and the magnetic field returns to normal.

Another feature of the solar wind-magnetosphere interaction is a buildup of particles in the tail and the generation of large, crosstail electric currents. The particle buildup culminates with an explosive injection of particles towards Earth where they are guided by the magnetic field into the regions near the poles. Excitation of the atoms in the upper atmosphere by these particles results in the emission of light, which we see as the aurora. In electrical terms, the injection of particles constitutes a deflection of the crosstail current into the ionosphere where it produces an intense, east-west current, called the 'auroral alectrojet', around the night side of the Earth. The magnetic disturbances observed at high latitudes, called 'substorms', are the magnetic fields of this auroral electrojet, and are the principal cause of power-system problems (Fig. 3).

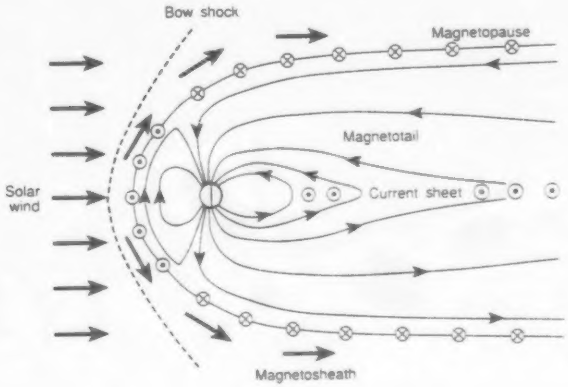


*Figure 1. The Earth's magnetosphere (*from *Hughes, 1995; reprinted with the permission of Cambridge University Press).*

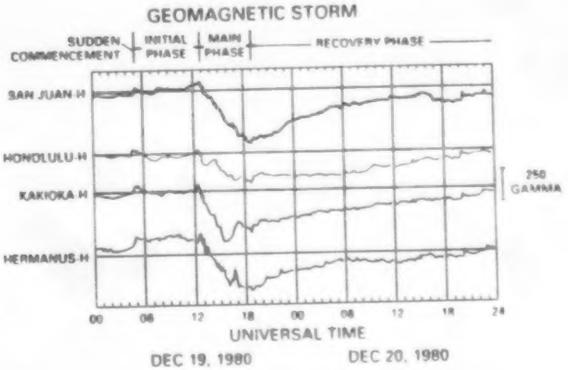


***Figure 2.** Variations of the northward magnetic field observed at four low-latitude locations around the Earth during a magnetic storm (*from *Campbell, 1997; reprinted with the permission of Cambridge University Press).*

GEOMAGNETIC CLIMATOLOGY

Regular measurements of the magnetic field started early in the nineteenth century and the first magnetic observatory in Canada was established in Toronto, Ontario, in 1839. The earliest observations involved manually noting the deflections of a magnetized needle. However, as early as 1847, photographic techniques were being used to automatically record magnetic-field variations. Photographic instruments were the principal means of recording at magnetic observatories for over a century and, in some countries, are still in use. It was only in the 1950s that electronic sensors came into use and only since the 1970s that digital recording has been used.

The magnetic recordings, called 'magnetograms', could be examined to learn about the detailed characteristics of magnetic-field variations, but there was also an interest in studying the overall distribution and occurrence of geomagnetic disturbances. For this it was necessary to find a measure of the strength of the magnetic disturbance, which could easily be determined from the magnetograms. The measure chosen was the range of variations in a specified time interval; to provide continuity, this is still the basis for the magnetic indices used today.

Magnetic indices

At a number of magnetic observatories around the world, the range of magnetic field variations in three-hourly intervals is routinely measured. These range measurement are converted to a 'K index' value from 0 (quiet) to 9 (major storm) that represents the size of the magnetic disturbance (*see* Fig. 4). These local K values are then combined to provide a global value of the size of the magnetic disturbance, called 'Kp', where p indicates that this is a planetary index; Kp values are available from 1932 to the present.

Mayaud (1973) measured the ranges from two magnetic observatories, one in the United Kingdom and one in Australia, to provide a measure of magnetic disturbances back to 1868. These were combined to give the 'aa index' of magnetic activity. To extend the record, Nevanlinna and Kataja (1993) derived aa values back to1844 using recordings made at the Helsinki magnetic observatory. With this we have been able to compile a record of geomagnetic activity over the entire period for which electrical systems have been in operation (Fig. 5).

SOLAR CYCLE VARIATIONS

The most fundamental factor affecting the occurrence of geomagnetic disturbances is their link to activity on the Sun. This was first established Sabine (1852) who compared magnetic

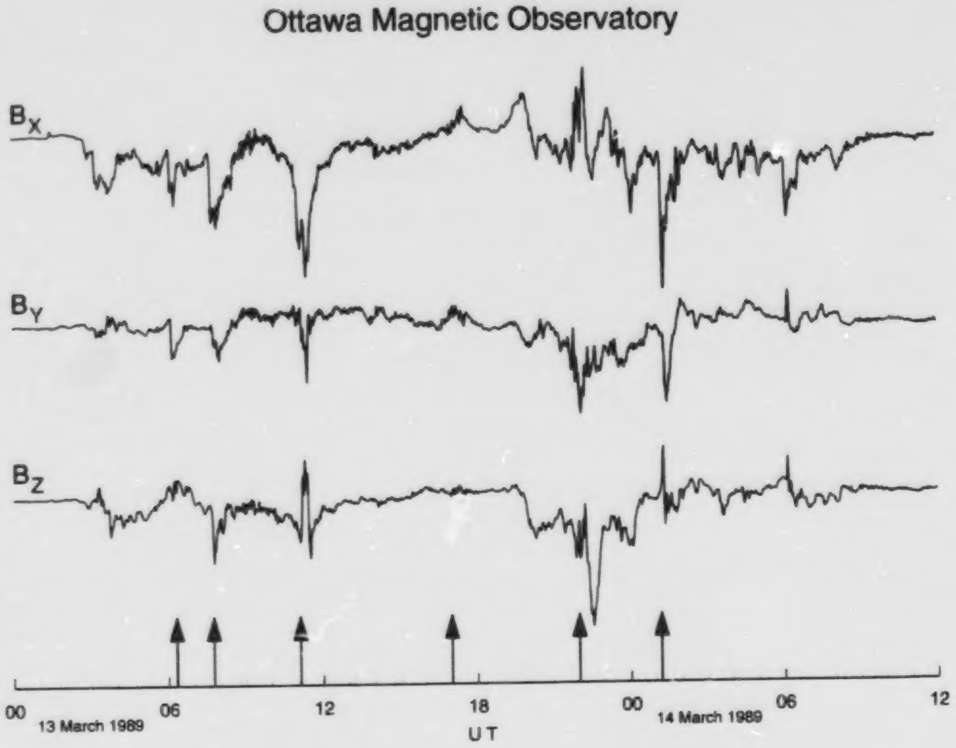


Figure 3. *Magnetic-field variations in the northward (B_x), eastward (B_y), and downward (B_z) directions recorded at the Ottawa Magnetic Observatory on 13-14 March 1989. The arrows show times of power system problems.*

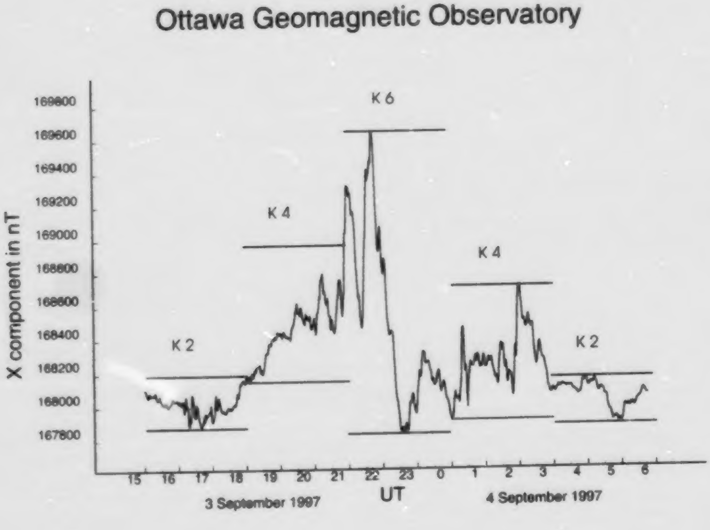


Figure 4.

Magnetic-field variations at Ottawa and corresponding K values.

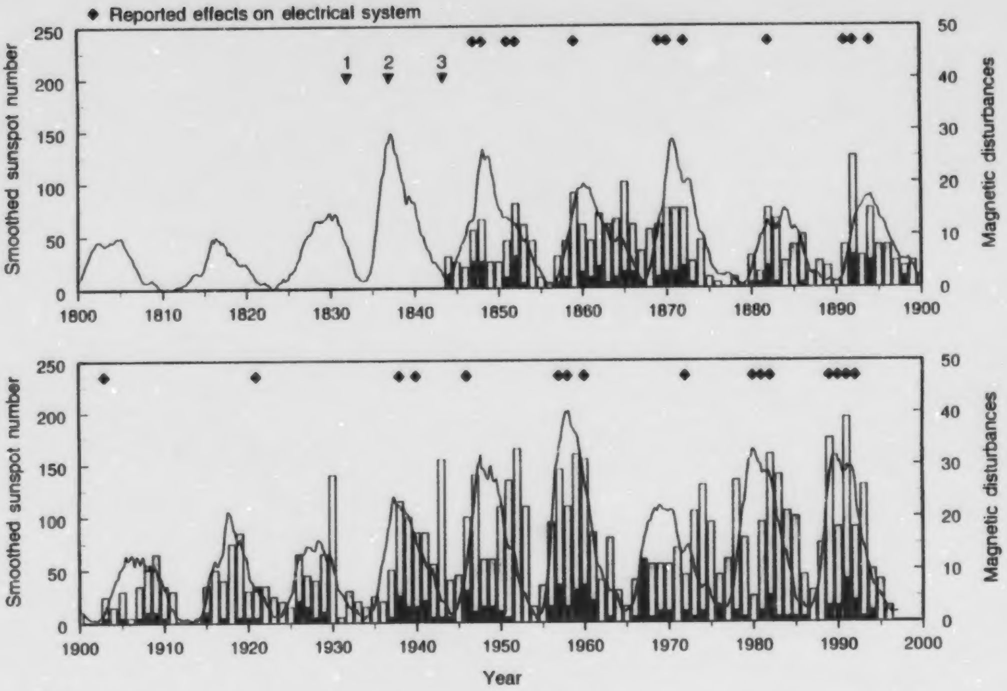


Figure 5. *Solar and geomagnetic activity and the occasions when this has disrupted the operation of long electrical systems on Earth. Grey bars are the number of magnetic disturbances with aa*>60 nT and black bars are the number of magnetic distrubances with aa*>120 nT. The solid line shows the smoothed sunspot number. Black diamonds represent the occurrence of geomagnetic effects affecting the operation of electrical systems. The development of long electrical systems is marked by 1) Faraday's discovery of electromagnetic induction in 1832; 2) Morse's and Wheatstone's patents for the telegraph in 1837; and 3) the perfection of Daniell's and Grove's batteries in 1843, that allowed the telegraph to be operated to significant distances* (from *Boteler et al., 1998*).

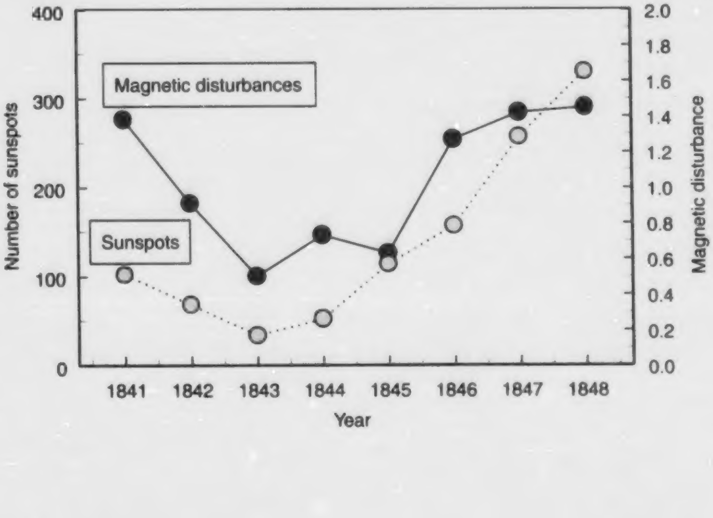


Figure 6.

Relationship between magnetic activity at Toronto (expressed as a ratio relative to the average intensity of the disturbance) and the number of sunspot groups (after *Sabine, 1852*).

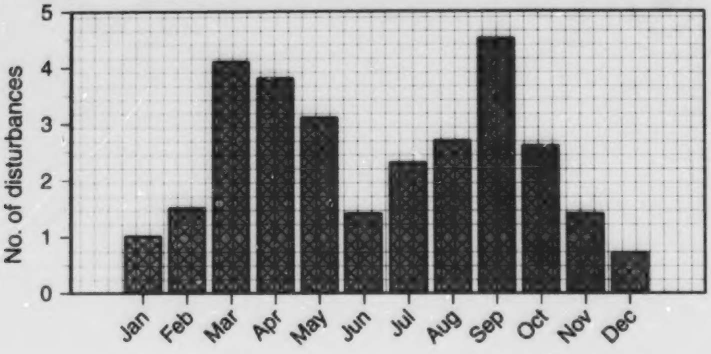


Figure 7.

Average monthly occurrences of magnetic disturbances with Kp of 7, 8, and 9.

disturbances recorded at the Toronto magnetic observatory with the number of sunspot groups seen each year on the surface of the Sun (Fig. 6). Sunspot numbers follow an approximately 11-year cycle. Burbank (1905) mentioned that the years 1848, 1859–1860, 1872, 1883, and 1903 were characterized by particularly severe disturbances to telegraph operation, and commented that the major disturbance on 30–31 October 1903 occurred almost exactly 55 years after the heavy disturbances of 27–28 October 1848. Figure 5 shows the smoothed sunspot cycle and the occurrence of moderate (aa*>60nT) and severe (aa*>120nT) magnetic disturbances, where aa* is a 24-hour running mean of the three-hourly aa magnetic index values. Occurrences of geomagnetic effects on the operation of electrical systems are described in Appendix A.

Seasonal variation

Geomagnetic activity not only follow the sunspot cycle, but it also varies seasonally because of changes in the energy transfer between the solar wind and the magnetosphere. The efficiency of this energy transfer varies with the change in orientation of the Earth as it orbits the Sun, producing seasonal variations in the occurrence of large magnetic disturbances, with a maximum in spring and autumn. Not many large magnetic disturbances occur each year, so the seasonal variation is not always apparent. However, by averaging the occurrence of magnetic disturbances with Kp of 7, 8, or 9 over 55 years, the underlying seasonal modulation can be readily seen (Fig. 7).

Diurnal variations

The occurrence of magnetic-field fluctuations also shows a distinct variation with time of day. However, this 'local time' variation in activity is not the same at all latitudes. To better understand the change in occurrence with latitude, it is useful to plot the results as a contour plot viewed from above the pole. Different local times correspond to fixed positions in space through which an observatory is carried by the Earth's rotation. Thus the variations in magnetic activity with local time correspond to the variations in activity with position around the circle of latitude of the observing site. This has been examined in detail in a study conducted by the Geological Survey of Canada in partnership with the Canadian power industry. Statistics on the peak rate of change of the magnetic field were compiled using 16 years of data from the Canadian Magnetic Observatory Network, which extends from middle latitudes to near the magnetic pole. Figure 8 shows the occurrence of hourly maximum rate of change values greater than 60 nT/minute.

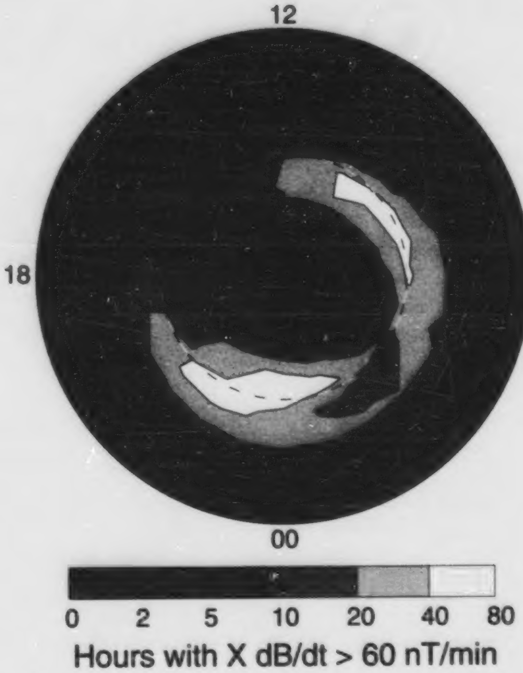


Figure 8. *Average yearly occurrences of hours with a rate of change of the magnetic field greater than 60 nT/ minute as a function of time of day (shown by the numbers around the edge of the plot). The edge of the plot is at geomagnetic latitude 50°. Inner circles show geomagnetic latitudes 60°, 70°, 80°* (from *Boteler et al., 1997).*

These plots clearly show that peak magnetic activity occurs around a roughly circular band extending from the evening sector to the noon sector. This activity region extends to lower latitudes on the night side of Earth, but is confined to higher latitudes on the day side. Thus, a middle latitude

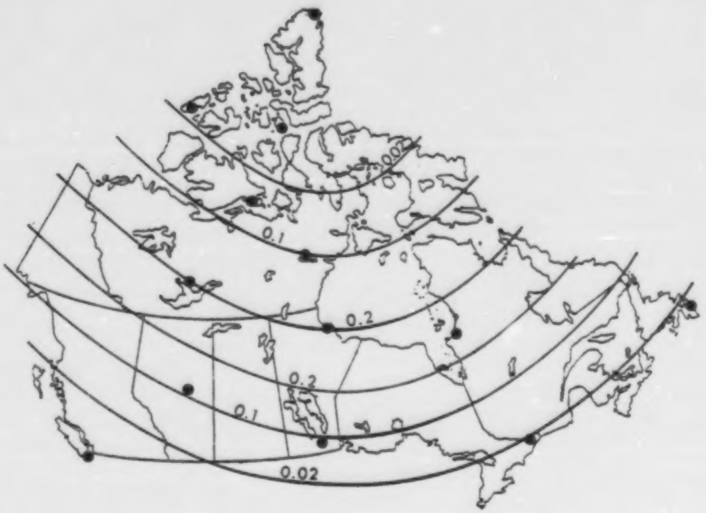


Figure 9.

Map of average geomagnetic activity, expressed as the percentage probability of occurrence of an hour during which dB/dt will equal or exceed 300 nT/min (from *Coles and Boteler, 1992).*

location is carried by the Earth's rotation through the disturbance region on the night side, but misses the higher latitude disturbance region on the day side. This produces the disturbance pattern seen at Ottawa, Ontario. At a location at a higher latitude, such as Churchill in northern Manitoba, the disturbed region at about latitude 70° N extends from the evening sector to the noon sector. A high-latitude site, such as Resolute, passes poleward of the night side disturbance, but is strongly affected by the disturbance region extending up to high latitudes on the day side.

Latitudinal variations

There is a distinct change with latitude in the occurrence of large magnetic variations. The occurrence of hours with high magnetic rates of change increases considerably as one goes north to the auroral regions and then shows a decrease as one moves from the auroral region to the polar cap. This relates to the proximity to the auroral electrojets that are the principal cause of the magnetic disturbances in Canada. Figure 9 shows the average probability of occurrence of hours with a magnetic rate of change that will equal or exceed 300 nT/minute, which represents a level of magnetic disturbance likely to have a significant effect on technological systems. The band across the country with the highest probabilities shows where power systems, pipelines, and cables will be most affected by geomagnetic effects. The maximum percentage probability of 0.2% in Figure 9 corresponds to an average occurrence of severe magnetic disturbances of slightly less than once a year.

HAZARD ASSESSMENT

Assessing the geomagnetic hazard to electrical systems involves calculating the electric fields produced at the Earth's surface during geomagnetic disturbances. The size of the electric fields depends on the rate of change (frequency content) of the magnetic-field variations and the electrical resistivity of the Earth. Electric fields calculated from this information can then be used as the input to system models to determine the response of power systems, pipelines or, telephone cables.

Magnetic field spectra

An extensive analysis of magnetic disturbances has been made as part of a study of geomagnetic hazard jointly funded by the Geological Survey of Canada and the Canadian Electrical Association. This study used data from the Canadian Magnetic Observatory Network operated by the Geological Survey of Canada (Fig. 10). At each observatory, the three components of the Earth's magnetic field are recorded once a minute and the data are transmitted via satellite to the Geomagnetic Laboratory in Ottawa. These data were used to determine the frequency spectrum of the magnetic field variations during different levels of magnetic activity.

All the results show the dominant features of the magnetic field spectra: a decrease in amplitude with increasing frequency and an increase in spectral amplitude (across the entire frequency band) with the level of magnetic activity, as given by Kp. A comparison of the spectra from different observatories also shows an increase in spectral levels with proximity to the auroral electrojets.

Resistivity structure

The electric fields produced by geomagnetic disturbances penetrate deep into the Earth, and the resistivity at these depths influences the size of the electric field seen at the surface. Therefore information is needed about the resistivity of the Earth down to many hundreds of kilometres below the surface. As part of the study mentioned above, published electrical survey results have been reviewed and compiled into a database of resistivity information. This has been used to derive layered-Earth models of the conductivity of different regions of Canada.

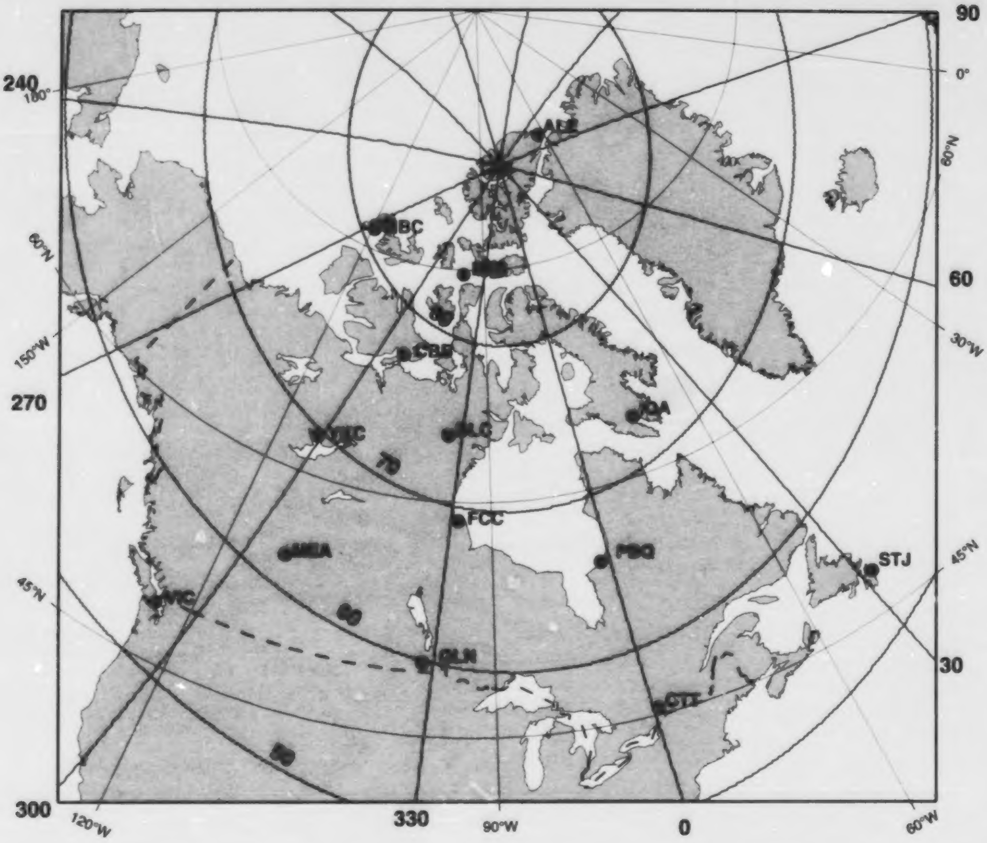


Figure 10. Location of Canadian geomagnetic observatories. The heavy lines are invariant geomagnetic co-ordinates (from *Boteler et al., 1997). ALE = Alert; BLC = Baker Lake; CBB = Cambridge Bay; FCC = Churchill; GLN = Glenlea; IQA = Iqaluit; MBC = Mould Bay; MEA = Meanook; OTT = Ottawa; PBQ = Poste-de-la-Baleine; RES = Resolute; STJ = St. John's; VIC = Victoria; YKC = Yellowknife*

The Canadian Shield represents the ancient heart of the North American continent and is composed of Precambrian rocks. These igneous and metamorphic rocks tend to have very high resistivity, but they also have some localized conductivity anomalies due to graphite, shear zones, etc. The shield rocks make up the surface of most of Labrador, Quebec, and Ontario, although they are overlain by a layer of sedimentary rocks under and southwest of Hudson Bay.

The Prairie Provinces, Alberta, Saskatchewan, and Manitoba, are in the Central Stable Region where many of the ancient shield rocks are covered by layers of Paleozoic sedimentary rocks. The sedimentary rocks are typically less than 2 km thick, but may be as much as 3 km thick in some basins. They have a higher water content and therefore lower resistivity than the rocks of the shield.

On the edges of the continent, episodes of mountain building and subsequent erosion have resulted in more complex rock formations. In the east, the opening and closing of the Atlantic Ocean basin produced the folded rocks of the Appalachian system that make up most of Nova Scotia, New Brunswick, and Newfoundland. On the west coast, the mountain ranges are produced by the collision of the North American Plate with the Pacific Plate. The subduction of the Pacific Plate below the North American continent also gives rise to volcanic activity that has modified the surface geology of the area.

In coastal regions, the electromagnetic response is strongly affected by the presence of the conducting seawater. Electric currents are induced in both the shallow waters of the continental shelf and the deeper ocean waters farther out. On the west coast, the continental margin is just beyond the Queen Charlotte Islands, about 150 km from the mainland. On the east coast, the continental shelf extends about 250 km from Nova Scotia and about 500 km beyond Newfoundland.

Calculation of electric fields

Using the conductivity information, surface impedance functions were calculated for each region of Canada. The magnetic-field spectra were then calculated, and the magnetic spectral values at each frequency were multiplied by the corresponding surface impedance value to give the spectral component of the electric field. These calculations were performed for each three-hour interval in 1991 for both northward and eastward components of the electric field, producing a total of 5840 spectra at each site. These electric-field spectra were then sorted and averaged according to time interval and Kp value.

To determine the typical magnitudes of the electric fields produced during different levels of magnetic activity, an inverse fast Fourier transform of the individual electric field spectra was used to calculate the time variations of the electric field. As with the spectral calculations, these calculations were made for the northward and eastward components at each observatory and for each three-hour interval in 1991. This gave sufficient information to fully define the magnitude and direction of the electric field at any time throughout 1991.

Occurrence of large electric fields

For each region, the peak magnitude of the electric field variations in each three-hour interval was also related to the corresponding values of Kp. Figure 11 shows the results for central Canada. All the results show a general increase in electric field magnitude with increasing Kp.

Combining these results with the occurrence of Kp values derived from the geomagnetic climatology shows that, for most parts of the country, an electric field of approximately 1 V/km can be expected, on average, once a year. In practice, most of these events can be expected to occur in the three or four years near the peak of the solar cycle. These results also show that electric-field magnitudes of about 3 V/km can be expected once every 10 years, i.e. once every solar cycle.

EFFECTS ON POWER SYSTEMS

The magnetic storm of 24 March 1940 produced the first reported effects on power systems. Many utilities in the northern United States reported problems ranging from voltage dips and reactive power swings to tripping of transformers. Problems occurred again during the next period of major storms, including incidents on 22 September 1957 and 11 February 1958 when a blackout occurre in the Toronto area. Additional effects were noted by power utilities across North America during the magnetic storm of 4 August 1972. Manitoba Hydro, Hydro-Québec, and the Newfoundland and Labrador Power Commission reported voltage fluctuations and tripping of equipment; quasi-DC currents of over 100 amperes were measured in transformers ground connections at substations at Corner Brook, Newfoundland, at La Verendrye, Quebec, and at Grand Rapids, Manitoba.

One of the worst geomagnetic storms recorded this century occurred on 13–14 March 1989. The most serious effect of the storm occurred at 02:45 EST on 13 March when widespread geomagnetically induced currents caused the saturation of transformers on the Hydro-Québec power system. The resulting harmonics caused the tripping of several static reactive power compensators, especially at Chibougamau and La Vérendrye substations. This led to voltage oscillations and power swings that caused the lines from James Bay to trup and the system to collapse. The Northeast Power Coordinating Council and the Mid-Atlantic Area Council power pools, which serve the entire northeastern United States from New England to Washington, D.C., were nearly involved in a cascading system collapse. Public Service Electric & Gas Co. did experience a series of transformer failures at its Salem nuclear generating station in New Jersey that resulted in extensive equipment damage.

High-voltage power distribution is provided by a network of three-phase transmission lines connected between Y-configuration transformers as shown in Figure 12. During normal operation, the different phase currents in each winding of the transformer add up to zero at the centre (neutral)

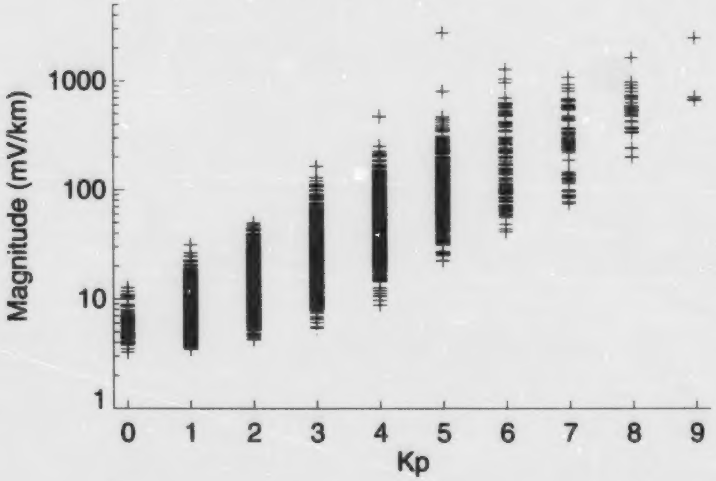


Figure 11.

Peak magnitudes of the electric field in central Canada for different values of the magnetic activity index Kp (derived using data from the Glenlea Magnetic Observatory) (from *Boteler et al., 1997).*

point of the Y, and no AC current flows along the connection to ground. However, when the AC supply is unbalanced or there are fault conditions, such as a broken conductor or a lightning strike on the line, the currents do not cancel at the neutral point and the connection to ground is needed to provide a safe discharge path. The presence of this neutral-ground connection allows geomagnetically induced currents in the transmission lines to flow to ground through the transformer windings, which causes a variety of problems.

Geomagnetically induced currents are driven by the electric fields produced during a geomagnetic disturbance. The sequence of phenomena is shown in Figure 13. Because of

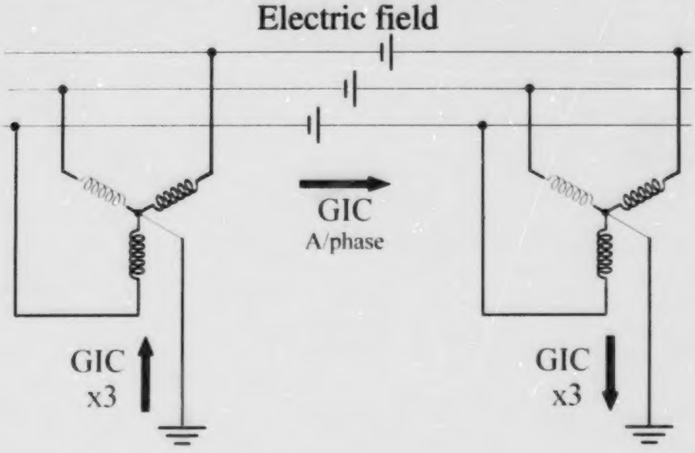


Figure 12.

Geomagnetic induction in a three-phase network. Geomagnetically induced currents (GIC) can flow through the neutral-ground connection into the transformer windings and along the transmission lines and back to ground through the transformers at another substation.

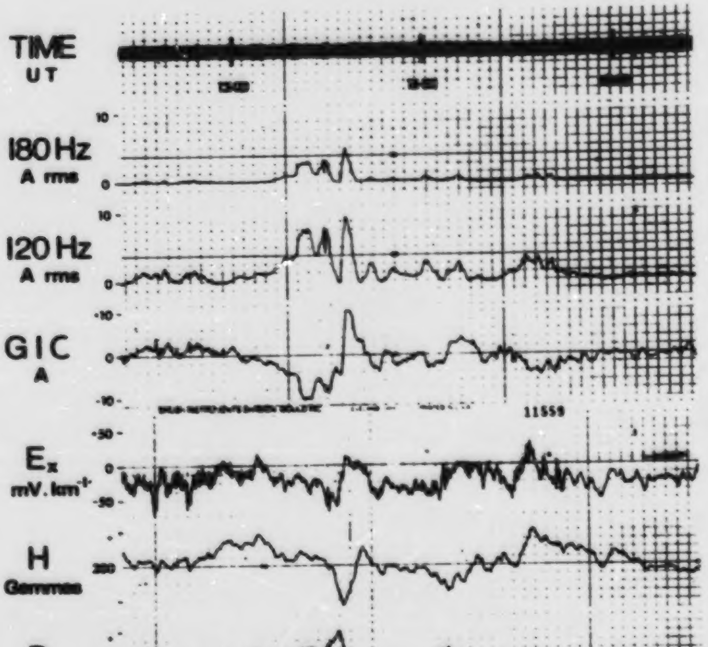


Figure 13.

Geomagnetic field variations (H, D, and Z) induce electric fields (E_X) that drive geomagnetically induced currents (GIC) through the power system. This causes partial saturation of transformers resulting in a distorted AC waveform with increased 2nd (120 Hz), 3rd (180 Hz), and other harmonics (observations made at Prince George on the BC Hydro system) (from *Boteler, 1991*).

their low frequency compared to the 60 Hz AC frequency, geomagnetically induced currents appear to a transformer as a slowly varying DC current. The geomagnetically induced current flowing through the transformer winding produces extra magnetization, which, during the half-cycles when the AC magnetization is in the same direction, can saturate the core of the transformer. This results in a very spiky AC waveform with increased harmonic levels that can cause relays and other equipment on the system to malfunction and lead to problems ranging from tripping of individual lines to collapse of the whole system.

Saturation of the transformer core causes increased transformer heating because of the extra electrical eddy currents produced in the transformer core and structural supports. Because of the large thermal mass of a high-voltage power transformer, this heating produces a negligible change in the overall transformer temperature. However, localized hot spots can occur and can damage the transformer windings. As an example of this, Figure 14 shows that a 60 A spike in the geomagnetically induced current flowing through the transformer caused the temperature of a hot spot on the tank wall to exceed 150°C, whereas no noticeable change occurred in the temperature of the transformer oil. During the March 1989 magnetic storm, a 500kV/138kV transformer at Meadow Brook substation near Winchester, Virginia, was taken out of service. Subsequent inspection showed four areas of discoloured paint on the tank that coincided exactly with the areas of maximum stray flux density if the core was saturated. The tank-wall temperature was estimated to have reached 400°C.

The financial consequences of a transformer failure go far beyond the replacement cost of the transformer. The transformer at the Salem nuclear generating station, in New Jersey, that burnt out during the 13 March 1989 magnetic disturbance cost several million dollars to replace. Such expensive units are usually built to order and the delivery time is normally about one year. Unusually, in this case, a replacement transformer was available and delivery and installation took only six weeks. Even so, having the transformer out of service limited the power that could be delivered from the Salem generating station, and the purchase of replacement power from neighbouring utilities cost about $17 million, far more than the cost of the transformer.

A novel feature of the effect of geomagnetically induced currents on power systems is that problems occur simultaneously on many systems, unlike other types of power-system problems, e.g. lightning strikes or equipment failures, which are more localized. The interconnection of modern power systems is designed to provide safeguards against localized failures, but may contribute to increased vulnerability to geomagnetically induced currents. Systems that had few problems during the last solar cycle are now operating

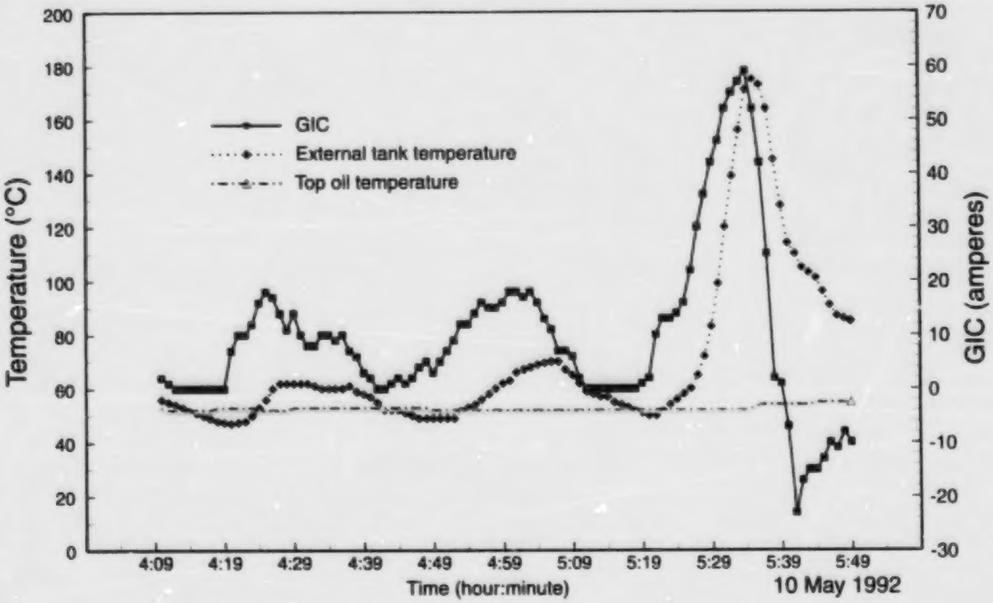


Figure 14. *Geomagnetically induced current (GIC), tank temperature, and oil temperature measured on a transformer on the Allegheny Power system (redrawn from results supplied by P. Gattens and reported in Kappenman, 1994). Geomagnetically induced currents do not affect the overall transformer temperature (shown by the oil temperature), but can produce hot spots on the external tank wall (*from *Kappenman, 1994).*

closer to their limits, and electric utility companies are concerned about the threat posed to the systems by geomagnetic disturbances during the peak of the next solar cycle.

EFFECTS ON PIPELINES

Prevention of pipeline corrosion requires that the electrical potential of the pipeline relative to the surrounding soil be maintained within a narrow range. Periodic surveys of the pipeline are carried out to ensure that the pipeline potential is correct. On occasions, during these surveys, pipeline engineers have noticed that the pipeline potential can fluctuate wildly. Such occasions coincide with disturbances in the Earth's magnetic field and are due to 'telluric' currents produced by the changing magnetic field. These telluric currents are of greatest concern for pipelines at higher latitudes where the magnetic disturbances are most intense, but have also been observed at lower latitudes, such as in Africa and Australia. As well as interfering with pipeline-potential surveys, telluric currents are of concern to pipeline engineers because they may contribute to pipeline corrosion. A comprehensive understanding of telluric currents has proven difficult to obtain because these currents are sporadic and vary both in time and with position along the pipeline. Single-instrument measurements on pipelines have shown different readings at different times and at different places, and the few simultaneous measurements made have shown widely different, even opposing, variations. Although engineers have been able to identify the characteristics of the disturbances on particular pipelines, until recently, no theoretical framework has existed to explain those observations or to make predictions about the expected effects of telluric currents.

An international study, led by the Geological Survey of Canada and involving pipeline companies from Canada, Norway, Sweden, and Finland, is underway to improve our understanding of the effects of telluric currents on pipelines. In this study, distributed-source transmission-line theory is used to simulate the electrical response of pipelines to telluric currents. This work shows that the electrical response of a pipeline depends on the resistance along the pipe and the resistance of the pipeline coating. These parameters determine an 'adjustment distance', which is a measure of how far telluric-current effects spread out along a pipeline. The pipe-to-soil potential profile produced by telluric currents depends on whether the pipeline is long or short in relation to its adjustment distance. Distributed-source transmission-line modelling shows that for a long, straight pipeline, the maximum pipe-to-soil potential variations occur at the ends of the pipeline and are independent of pipeline length. The theory also shows that potential variations at opposite ends of the

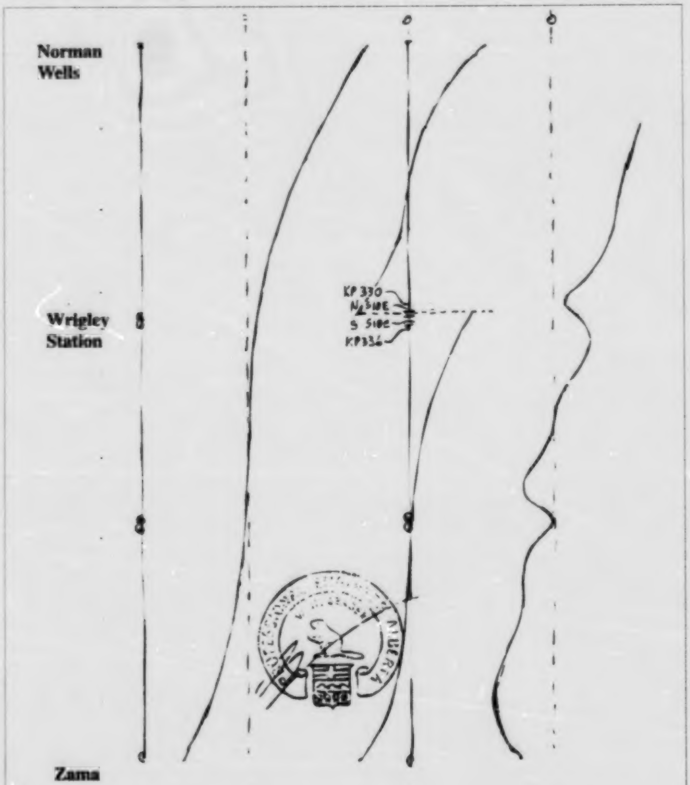


Figure 15.

*Telluric pipeline-potential profile obtained from observations on the Norman Wells–Zama pipeline. The curves show the profile for **a**) the whole line connected throughout, **b**) the line with an insulating flange at Kilometre Point 335, and **c**) a connected line, including the effects of station grounding* (from *Boteler and Seager, 1998; reprinted with the permission of the National Association of Corrosion Engineers).*

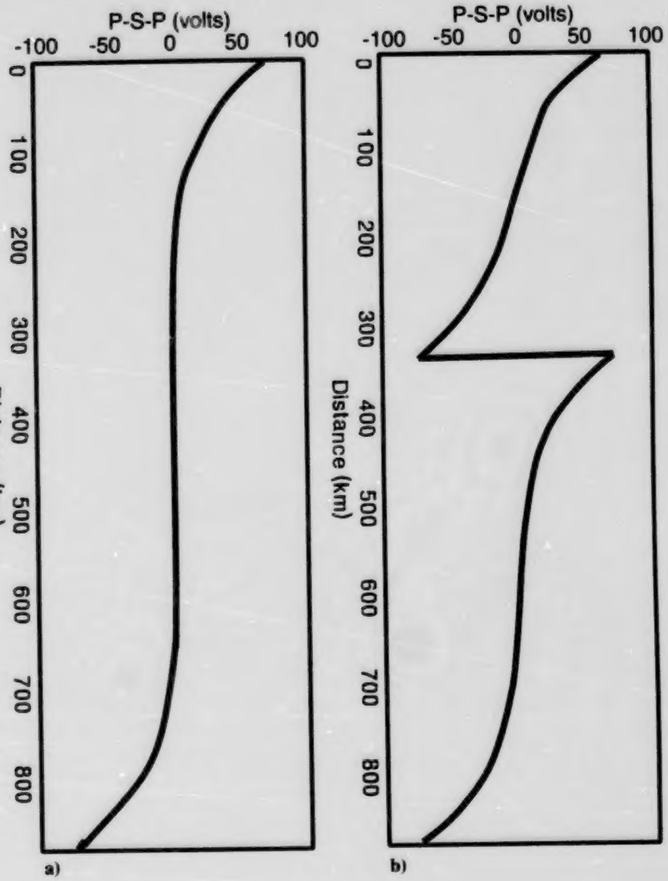


Figure 16.

Results of distributed-source transmission-line modelling for the Norman Wells–Zama pipeline: ***a****) assuming a straight, electrically continuous pipeline;* ***b****) for a pipeline with an insulating flange at Kilometre Point 335* (from *Boteler and Seager, 1998; reprinted with the permission of the National Association of Corrosion Engineers).*

pipeline are out of phase. Pipeline potentials fall off exponentially with distance from either end and cross through zero in the middle of the pipeline.

Measurements on long pipelines have been difficult to carry out and normally require many simultaneous recordings of potential variations to determine the pipeline reponse to telluric currents. However, by leapfrogging two potential recorders by helicopter along the Norman Wells–Zama pipeline in northwestern Canada, W.H. Seager was able to examine the phase relationship of the telluric variations on the pipeline (Boteler and Seager, 1998). Also, by comparing pipeline-potential recordings to magnetic recordings from the GSC's Yellowknife magnetic observatory, Seager was able to construct a schematic picture of how the amplitude of telluric fluctuations varied along the pipe (Fig. 15). Additional tests were conducted at Wrigley Station (Kilometre Point 335) where an insulating flange, normally shorted by a bond wire, was located. When the bond was opened, the telluric potential increased substantially on both sides and a phase reversal of the telluric potentials on opposite sides of the insulating flange was noted.

Model calculations of the potential profile along the Norman Wells–Zama pipeline were made for comparison with the observations (Boteler and Seager, 1998). Figure 16 shows the potential profile for an electric field of 1 V/km parallel to the pipeline. It also shows the calculated potential profile for the pipeline with an insulating flange at Kilometre Point 335. The calculated potential profile (Fig. 16) agrees very well with the potential profile derived from observations (Fig. 15). They both show the characteristic S-shaped profile, with potential variations at one end of the pipeline being opposite in sign to those at the other end, of an 'electrically long' pipeline. Both the model results and the observations show that the maximum voltages occur at the ends of the pipeline and that large voltage variations are produced on either side of an insulating flange.

One suggestion for countering telluric currents has been to place insulating flanges in the pipe to interrupt the current flow. For an electrically long pipeline, observations and pipeline modelling show that an insulating flange just creates another site where pipeline potential can build up and the telluric-current flow along the pipe is forced to ground. Thus this method would actually lead to an increased risk of corrosion.

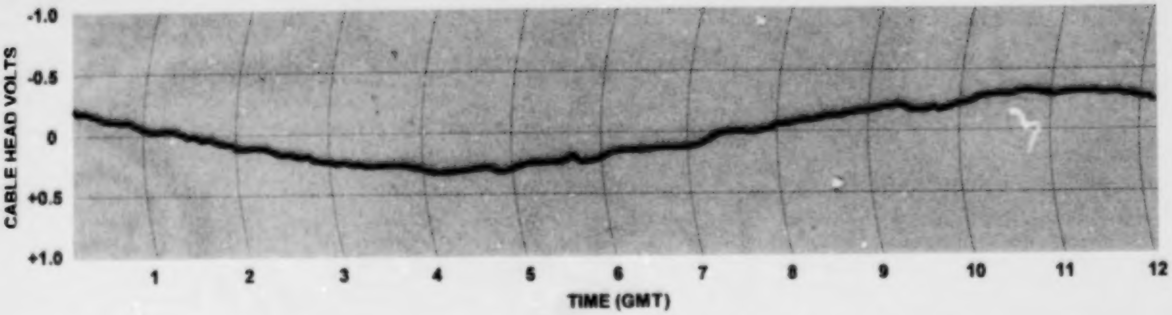


***Figure 17.** Cable voltage variations on the Port Kail– Donaghadee submarine cable produced by tidal water movements through the Earth's magnetic field in the Irish Sea* (from *Axe, 1968*).

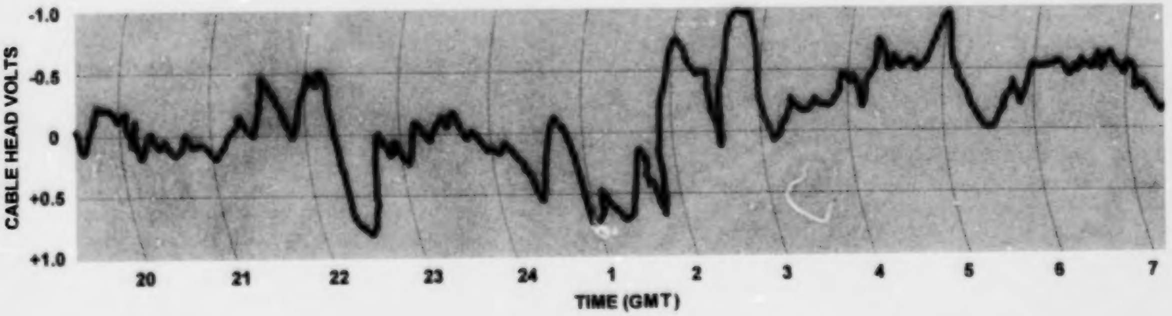


***Figure 18.** Cable voltage fluctuations on the Port Kail– Donaghadee submarine cable produced by magnetic disturbances* (modified from *Axe, 1968*).

There is growing recognition that the design of corrosion-prevention systems needs to take into account the 'geomagnetic dimension'. The use of higher resistance coatings on modern pipelines has meant that potentials produced by telluric currents are larger than on older pipelines. Distributed-source transmission-line theory now provides a means of modelling the pipeline potentials produced by telluric currents. The modelling provides information on the pipe-to-soil potentials to be expected at different locations along a pipeline and provides a theoretical framework that helps explain many observations. This work will eventually lead to the capacity to simulate telluric currents in pipeline networks and to predesign cathodic protection systems that can also counteract any adverse telluric effect.

EFFECTS ON COMMUNICATION CABLES

The telegraph was the first man-made system to involve the use of long electrical conductors. At times, telegraph operators were troubled by fluctuations on the system caused by telluric currents flowing from the Earth. In the early days of the telegraph, a variety of methods were used to record the signal transmitted over the wires. Bain's 'chemical telegraph' (Prescott, 1866) used specially prepared paper that was sensitive to an electric current. The current from a stylus touching the paper caused a chemical reaction, which left a coloured mark on the paper. During the magnetic storm of 19 February 1852, the current increased so much that a "flame of fire" followed the pen and set fire to the prepared paper (Prescott, 1866, p. 318). A magnetic storm from 28 August to 2 September 1859 produced widespread effects on the telegraph system in Europe and North America. Widespread disturbances to telegraphic communication occurred in 1872, and on 31 October 1903, practically the world's entire telegraph system was upset.

On 24 March 1940, a major magnetic storm affected both the telegraph system and the power system in North America. Stetson (1947, p. 79) recounts how the disturbance disrupted the transmission of Easter telegrams and gives the following example of one message that was received on the teletype:

> GOVE AND EAETER GREETINGS FROM
> AGLWGRACZMGOPBWQQTPOZLQPVPMGXPZGXH
> VWYQPMRQPWVVZPVW"/%/WXQQVVWVPVMOZ
> QOQQQWQWVWQMVN.

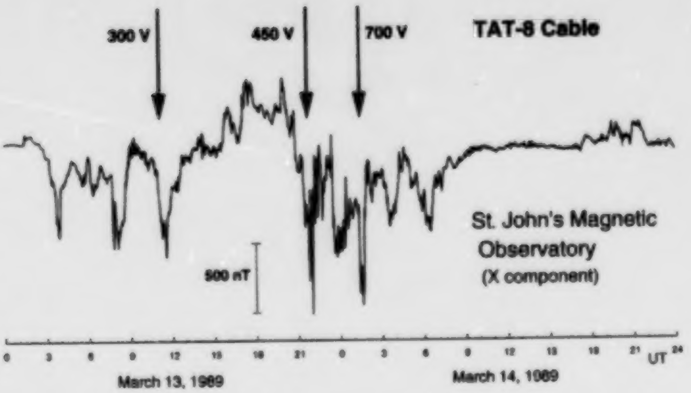


Figure 19.

Variation of the magnetic field at the St. John's Magnetic Observatory and peak voltage excursions (arrows) measured on the power supply for the fibre optic transatlantic telecommunication cable on 13–14 March 1989.

Telephone communications in the United States were also disrupted, the most serious effects occurring on the Eastern Seaboard and in North Dakota, Minnesota, Iowa, and Missouri. Newspaper headlines from the 1940s show that magnetic disturbances continued to cause problems to communication systems.

The introduction of coaxial cables increased the bandwidth of communication systems, but required the use of repeater amplifiers to compensate for signal losses due to cable resistance. To provide equal power to all repeaters along a cable, the repeaters are connected in series and powered by a constant-current supply at the ends of the cable. These power supplies have to be designed to maintain the current within narrow limits even when a large, geomagnetically induced voltage alternately adds to or subtracts from the cable's driving voltage. Thus, the power-feed equipment must have a wide output range and a fast enough response to cope with rapid swings in the induced voltage. On 4 August 1972, however, the magnetic disturbance was so severe that the induced electric field exceeded the operating limits of the L4 coaxial cable system in the midwestern United States, causing a complete system outage.

Submarine cables can be affected by both induced voltages caused by geomagnetic disturbances and the motional electromotive force produced by the movement of the conducting seawater through Earth's magnetic field. The motional electromotive force is largest when fast tidal flow through a channel is fast. Such tidal voltages have been detected on cables across the English Channel, the Irish Sea, and elsewhere. Water flow in one direction will produce a particular voltage, and when the water flow reverses with the change of the tide, the voltage changes polarity, producing a sinusoidal variation with an approximately 12 hour period, as shown in Figure 18. When magnetic storms occur, the voltages induced in the cable fluctuate more rapidly than the tidal voltages and can completely obscure the tidal effects (Fig. 19).

On short cables, repeaters are not used and the tidal voltage adds directly to the signal transmitted through the cable. This produced an interesting effect on the transmission of a music channel's signal from Britain to Northern Ireland, across the Irish Sea. After complaints about poor audio quality at various times, it was found that attenuation of the low-frequency components varied with the state of the tide. Subsequent investigations showed that, at times of maximum water flow, the tidal voltage produced an extra current in the cable that saturated the audio-frequency transformers at the ends of the cable. This effect was corrected by fitting a large series capacitor in the circuit to block the nearly DC currents produced by the tidal voltage.

During the magnetic storm of 10 February 1958, transatlantic communication from Clarenville, Newfoundland, to Oban, Scotland, alternated between loud squawks and faint whispers as the geomagnetically induced voltage acted with or against the cable supply voltage. Transoceanic and other long cables use repeaters to amplify the signal as it proceeds along the cable. These submerged repeaters are connected in series with the cable's centre conductor and are powered by a direct current supplied from the terminal stations at the ends of the cable. The power-feed equipment operates in constant-current mode and the extraneous voltages produced by magnetic disturbances cause a change in the output voltage of the equipment. Thus the size of the induced voltages and their frequency content have to be taken into account in determining the maximum output voltage and the speed of regulation required for the power-feed equipment. In spite of these precautions, in severe cases the total induced voltage in the cable can become so large that it exceeds the limits of the protection circuits on the cable power supply and the system automatically shuts down.

New submarine cables use optical fibres to carry the signals, but a conductor through the cable is still used to carry the power to the repeaters, so these systems remain susceptible to geomagnetic effects. At the time of the March 1989 storm, a new transatlantic telecommunications fibre-optic cable was in use. No disruption occurred, but large induced voltages were observed on the power-supply cables. The largest voltage excursions were observed at approximately 11:10 UT and 21:45 UT on 13 March and at 01:30 UT on 14 March. In Figure 157, magnetic recordings from the St. John's magnetic observatory in Newfoundland clearly show that peak cable voltage excursions coincide with the largest variations of the magnetic field. In future, because of improvements in fibre optics, cables may use fewer repeaters and require a lower driving voltage. However, downsizing the power-feed equipment without taking into account the induced voltages may leave these systems more vulnerable to geomagnetic effects.

REFERENCES

Acres Consulting Services Ltd.
1975: Study of the disruption of electric power systems by magnetic storms; Earth Physics Branch, Open File 77-19, 30 p.

Albertson, V.D. and Thorson, J.M.
1974: Power system disturbances during a K-8 geomagnetic storm: August 4, 1972; Institute of Electrical and Electronics Engineers, IEEE Transactions on Power Apparatus and Systems, v. PAS-93, p. 1025–1030.

Allen, J.
1994: Catalogue of magnetic disturbances using the aa* index; World Data Centre C-1, Boulder, Colorado.

Anderson, C.W., Lanzerotti, L.J., and Maclennan, C.G.
1974: Outage of the L-4 system and the geomagnetic disturbances of August 4, 1972; Bell System Technical Journal, v. 53, p. 1817–1837.

Angot, A.
1897: The Aurora Borealis; Appleton and Co., New York, 264 p.

Anonymous
1909: (untitled); Nature, v. 81, p. 523–524.

Aspnes, J.D., Merritt, R.P., and Akasofu, S.-I.
1981: Effects of geomagnetically induced current on electric power systems; The Northern Engineer, v. 13, no. 3, p. 34–38.

Axe, G.A.
1968: The effects of the earth's magnetism on submarine cable; Post Office Electrical Engineers Journal, v. 61, no. 1, p. 37–43.

Barlow, W.H.
1849: On the spontaneous electrical currents observed in wires of the electric telegraph; Royal Society of London, Philosophical Transactions, v. 139, p. 61–72.

Blais, G. and Metsa, P.
1993: Operating the Hydro-Québec grid under magnetic storm conditions since the storm of March 13, 1989; *in* Proceedings, Solar-Terrestrial Predictions Workshop, Ottawa, May 18–22, 1992, (ed.) J. Hruska, M.A. Shea, D.F. Smart, and G. Heckman; v. 1, p. 108–130.

Boteler, D.H.
1991: Predicting geomagnetic disturbances on power systems; EOS, Transactions of the American Geophysical Union, v. 72, no. 14, p. 159–160.

Boteler, D.H. and Jansen van Beek, G.
1993: Mapping the March 13, 1989 magnetic storm and its effects across N. America; *in* Proceedings, Solar-Terrestrial Predictions Workshop, Ottawa, May 18–22, 1992, (ed.) J. Hruska, M.A. Shea, D.F. Smart, and G. Heckman, v. 3, p. 57–70.

Boteler, D.H. and Seager, W.H.
1998: Telluric currents: a meeting of theory and observation; Corrosion, v. 54, p. 751–755.

Boteler, D.H. et al.
1997: Geomagnetically induced currents: geomagnetic hazard assessment, phase 2; Open File 3420, 140 p.

Boteler, D.H., Pirjola, R.J., and Nevanlinna, H.
1998: The effects of geomagnetic disturbances on electrical systems at the Earth's surface; Advances in Space Research, v. 22, p. 17–27.

Bozoki, B., Chano, S.R., Dvorak, L.L., Feero, W.E., Fenner, G., Guro, E.A., Henville, C.F., Ingleson, J.W., Mazumdar, S., McLaren, P.G., Mustaphi, K.K., Phillips, F.M., Rebbapragada, R.V., and Rockefeller, G.D.
1996: The effects of GIC on protective relaying; Institute of Electrical and Electronics Engineers, IEEE Transactions on Power Delivery, v. 11, no. 2, p. 725–739.

Burbank, J.E.
1905: Earth currents: and a proposed method for their investigation; Terrestrial Magnetism and Atmospheric Electricity, v. X, p. 23–49.

Campbell, W.H.
1997: Introduction to Geomagnetic Fields; Cambridge University Press, Cambridge, United Kingdom, 290 p.

Chapman, S. and Bartels, J.
1940: Geomagnetism; Oxford University Press, Oxford, United Kingdom, 1049 p.

Clement, K.T.
1860: Das grosse Nordlicht in der Nacht zum 29 August 1959 und die Telegraphenverwirrung in Nord-Amerika und Europe; Perthes-Besser & Mauke, Hamburg, 121 p.

Coles, R.L. And Boteler, D.H.
1992: Geomagnetic Hazard Assessment, Phase 1; Canadian Electrical Association Project 357T848A, Final Report, 81 p.

Davidson, W.F.
1940: The magnetic storm of March 24, 1940 — effects in the power system; Edison Electric Institute Bulletin, July 1940, p. 365–366, 374.

Elovaara, J., Lindblad, P., Viljanen, A., Mäkinen, T., Pirjola, R., Larsson, S., and Kielen, B.
1992: Geomagnetically induced currents in the Nordic power system and their effects on equipment, control, protection and operation; Proceedings, CIGRE (Conférence Internationale des Grand Réseaux Électriques à Haute Tension) Colloquium, 30 August–5 September, Paper 36-301, p. 1–10.

Finn, W.
1903: Influences of the sunspots upon electrical and magnetic forces of the earth; Scientific American Supplement no. 1457, p. 23351–23352.

Germaine, L.W.
1940: The magnetic storm of March 24, 1940 — effects in the communication system; Edison Electric Institute Bulletin, July 1940, p. 367–368.

Harang, L.
1951: The Aurorae; Chapman and Hall, London, United Kingdom, 163 p.

Hughes, W.J.
1995: The magnetopause, magnetotail, and magnetic reconnection; *in* Introduction to Space Physics, (ed.) M.G. Kivelson and C.T. Russell; Cambridge University Press, Cambridge, United Kingdom, p. 229.

Kappenman, J.G.
1994: Discussion of Meliopoulos et al.; Institute of Electrical and Electronics Engineers, IEEE Transactions on Power Delivery, v. 9, p. 194–207.
1996: Geomagnetic storms and their impact on power systems; Institute of Electrical and Electronics Engineers, Power Engineering Review, v. 16, p. 5.

Karsberg, A., Swedenborg, G., and Wyke, K.
1959: The influences of earth magnetic currents on telecommunication lines; Tele (English edition), Televerket, Stockholm, p. 1–21.

Lanzerotti, L.J. and Gregori, G.P.
1986: Telluric currents: the natural environment and interactions with man-made systems; *in* The Earth's Electrical Environment, (ed.) R. Roble and E.P. Krider; National Academy Press, Washington, DC, p. 232–257.

Lockyer, W.J.S.
1903: Magnetic storms, aurorae and solar phenomena; Nature, v. 69, no. 1775, p. 9–10.

Mayaud, P.N.
1973: A hundred years of geomagnetic data, 1968–1997, indices aa, storm sudden commencements; International Association of Geomagnetism and Aeronomy Bulletin 33, 252 p.

McNish, A.G.
1940: Magnetic storms; Edison Electric Institute Bulletin, July 1940, p. 361–364.

Nevanlinna, H. and Kataja, E.
1993: An extension of the geomagnetic activity index series *aa* for two solar cycles (1844–1868); Geophysical Research Letters, v. 20, p. 2703–2706.

Preece, W.H.
1894: "Earth Currents", Letter to the Editor; Nature, v. 49, no. 1276, p. 554.

Prescott, G.B.
1860: Observations made at Boston, Mass., and vicinity; *in* The great auroral exhibition of August 28th to September 4th, 1859, (ed.) B. Silliman; American Journal of Science and Arts, v. XXIX, p. 92–95.
1866: History, Theory and Practice of the Electric Telegraph (fourth edition); Ticknor and Fields, Boston, Massachussetts, 504 p.

Sabine, E.
1852: On periodical laws discoverable in the mean effects of the larger magnetic disturbances; Royal Society of London, Philosophical Transactions, p. 103–124.

Sanders, R.
1961: Effects of terrestrial electromagnetic storms on wireline communications; Institute of Radio Engineers, Transactions on Communications Systems (December), p. 367–377.

Silliman, B.
1860: The great auroral exhibition of August 28th to September 4th, 1859; American Journal of Science and Arts, v. XXVIII, p. 385-408, v. XXIX, p. 92–97, p. 249–266.

Silverman, S.M.
1995: Low altitude auroras: the storm of 25 September 1909; Journal of Atmospheric and Terrestrial Physics, v. 57, p. 673–685.

Slothower, J.C. and Albertson, V.D
1967: The effects of solar magnetic activity on electric power systems; Journal of Minnesota Academy of Science, v. 34, p. 94–100.

Stetson, H.T.
1947: Sunspots in Action: The Ronald Press Company, New York, 252 p.

Varley, C.F.
1873: Discussion of a few papers on earth currents; Journal of the Society of Telegraph Engineers, v. 2, p. 111–114.

Wallerius, A.
1982: Solen gav Sverige en strömstöt; Ny Teknik, v. 29, p. 3.

Winckler, J.R., Peterson, L. Hoffman, R., and Arnoldy, R.
1959: Auroral X-rays, cosmic rays, and related phenomena during the storm of Feb. 10–11, 1958; Journal of Geophysical Research, v. 64, p. 597.

APPENDIX A

150 years of geomagnetic effects

Many different man-made systems are affected during magnetic disturbances. Here, we have documented only the effects of geomagnetic disturbances on electrical systems at the Earth's surface, such as power systems, pipelines, and communication cables. The events listed here are incidents that can be related to a specific magnetic disturbance and for which a published source was found. Some failures, due to cumulative effects, have likely never been linked to geomagnetic effects. Such cumulative action has usually been associated with pipeline corrosion, but evidence now exists that geomagnetically induced currents may also produce cumulative stresses on generator step-up transformers (Kappenman, 1996).

The list of magnetic storms compiled by Allen (1994) using the aa index provides a useful guide as to when effects could be expected to have affected technological systems. Two notable magnetic disturbances occurred in 1946, one on 28–29 March, which had an aa* of 329 and is ninth in Allen's list of largest storms, and another on 22–23 September, which had an aa* of 295 and is the seventeenth largest storm ever recorded. Another notable event occurred on 18–19 September 1941, for which there are no known reports of affected systems. This is surprising because the magnetic storm had an aa* of 429 and is second only to the 13 March 1989 storm in the list of largest disturbances. More reports of system effects may be uncovered as further research is done on this topic.

1847

Barlow (1849, p. 61–62) was the first person to make systematic observations about spontaneous electrical currents observed in the wires of the electric telegraph.

> When the telegraph instruments are not working, the batteries are put out of circuit, and the wires remain with a simple earth connection at both extremities.
>
> It was in this condition of the wires that spontaneous currents were observed to arise in them, producing occasionally large deflections in the needles. These deflections were sometimes to the right and sometimes to the left; at times they changed rapidly from right to left, at others they continued in one direction for periods varying from a few minutes to one or more hours....
>
> My attention was strongly drawn to the subject by the constancy of these effects, when a circumstance occurred which imparted a new interest to the inquiry. On the evening of the 19th of March, 1847, a brilliant aurora was seen, and during the whole time of its remaining visible, strong alternating deflections occurred on all the instruments. Similar effects were observed also on the telegraphs on several other lines of railway.

1848

Heavy disturbances to the telegraph system occurred on 27–28 October (Burbank, 1905). Matteucci remarked on the coincidence of the aurora with interruptions in telegraphic communications, produced by telluric currents (Angot, 1897, p. 138).

On November 17, the appearance of the aurora coincided with effects on the electric telegraph between Florence and Pisa (Prescott, 1866).

1851

Prescott (1866, p. 318) reported "...a remarkable aurora, which took complete possession of all the telegraph lines in New England, and prevented any business from being transacted during its continuance."

1852

A brilliant auroral display was observed on 19 February. Associated with it were currents that burned through the chemical paper used with the Bain's chemical telegraph in the northeastern United States (Prescott, 1866).

1853

Varley (1873, p. 112) describes some of the early earth current observations: "In 1853 or 1854, the earth current on a length of two miles of wire between Telegraph Street and the Strand was so strong that is was impossible to work the line during its continuance."

1859

A magnetic storm from 28 August to 2 September produced widespread effects on the telegraph system in Europe and North America. Mr. O.S. Wood, Superintendent of the Canadian telegraph lines, described some of these effects (Prescot, 1866, p. 320):

> I never, in my experience of fifteen years in the working of telegraph lines, witnessed anything like the extraordinary effect of the aurora borealis, between Quebec and Father's Point, last night. The line was in most perfect order, and well-skilled operators worked incessantly from eight o'clock last evening till one o'clock this morning, to get over, in even a tolerably intelligible form, about four hundred words of the steamer Indian's report for the press; but at the latter hour, so completely were the wires under the influence of the aurora borealis, that it was found utterly impossible to communicate between the telegraph stations, and the line was closed for the night.

Problems were also reported by telegraph operators in New York, Washington, Philadelphia, Vermont, and Massachussetts (Clement, 1860; Prescott, 1860, 1866). Interesting observations were made in France by and are reproduced *in* Angot (1897, p. 141) as follows:

> At all the telegraphic stations in France the service was impeded during the whole of September 2 [1859], but especially at two periods of the day, from 4:30 A.M. to 9 A.M., and from noon to 3 P.M. These two periods were the same at all stations, and the greatest disturbances took place exactly at the same hours, at 7 A.M. and at 2 P.M. The longest wires always showed the greatest disturbances. The same day telluric currents were also observed in the greater part of the two hemispheres, in Switzerland, in Germany, in the British Isles, in North America, and throughout Australia.

Many reports from telegraph operators were collected *in* Silliman (1860), including the following:

> *Observations made at Boston, Mass., and its vicinity by* GEORGE B. PRESCOTT, *Telegraph Superintendent*
>
> ...The effects of the magnetic storm of August 28th, 1859, were apparent upon the wires during a considerable portion of Saturday evening, and during the entire day, Sunday. At 6 P.M. the line to New Bedford (60 miles in length, running a little west of south) could be worked only at intervals, although, of course, no signs of the Aurora Borealis were visible to the eye at that hour. The same was true of the wires running east through the state of Maine as well as those running north to Montreal. (Silliman, 1860, p. 92–93)
>
> On Friday, September 2d, 1859, upon commencing business at 8 o'clock A.M. it was found that all the wires running out of the office were so strongly affected by the auroral current as to prevent any business being done, except with great difficulty. (Silliman, 1860, p. 94)
>
> *5. Observations made at Philadelphia, communicated by* H. EMMONS THAYER, *Telegraph Superintendent*
>
> On the evening of Aug. 28th, about 8 o'clock, we lost current on all our four wires running from Philadelphia to New York, and we had strong circuit as if from a near ground connection; but there was no interruption on wires running south to Baltimore and Washington. At 9^h10^m the wires were relieved to a great extent from the influence of the Aurora, giving us our usual working current.
>
> On testing wires at 8 o'clock on the morning of Sept. 2d, I found two of our wires, those running via Camden and Amboy to New York, strongly under the influence of an Aurora. The effect was different from that of Aug. 28th. There was an intensity of current *which gave a severe shock* when testing, giving a reversed current, neutralizing our batteries, and destroying magnetism. On removing the batteries, we had a very strong circuit, giving powerful magnetism, but could not raise New York. On the line running from this city to Pittsburgh, the operator, Mr. Steacy, succeeded in transmitting a business message to Pittsburgh *wholly on the auroral current*. The current was changeable, suddenly disappearing and returning at intervals of from five to ten minutes. The signals were distinct and the conversion lasted four or five minutes, the operators exchanging remarks as to the singularity of the phenomenon. At 9. A.M. all the wires were relieved from the effects of the Aurora, and worked as well as usual. (Silliman, 1860, p. 96)
>
> *6. Observations made at Washington, D.C., by* FREDERICK W. ROYCE, *Telegraph operator*
>
> On the evening of Aug. 28th I had great difficulty in working the line to Richmond, Va. It seemed as if there was a storm at Richmond. I therefore abandoned that wire, and tried to work the northern wire, but met with the same difficulty. For five or ten minutes I would have no trouble, then the current would change, and become so weak that it could hardly be felt. It would then gradually change to a 'ground' so strong that I could not lift the magnet. The Aurora disappeared at a little after 10 o'clock, after which we had no difficulty. During the auroral display, I was calling Richmond, and had one hand on the iron plate. Happening to lean towards the sounder, which is against the wall, my forehead grazed a ground wire. Immediately I received a very severe electric shock, which stunned me for an instant. An old man who was sitting facing me, and but a few feet distant, said that he saw a spark of fire jump from my forehead to the sounder.
>
> *7. Observations made at Pittsburgh, Pa., communicated by* E.W. CULGAN, *Telegraph manager*
>
> During the Aurora of Aug. 28th the intensity of the current evolved from it varied very much, being at times no stronger than an ordinary battery, and then suddenly changing the poles of the magnets it would sweep through them, charging them to their utmost capacity, and compelling a cessation of work while it continued.
>
> On the morning of Sept.2d, at my request the Philadelphia operator detached his battery, mine being already off. We then worked with each other at intervals as long as the auroral current continued, which varied from thirty to ninety seconds. During these working intervals we exchanged messages with much satisfaction, and *we worked more steadily when the batteries were off than when they were attached*.
>
> On the night of Aug. 28th the batteries were attached, and on breaking the circuit there were seen not only sparks (that do not appear in the normal condition of a working line) but at intervals regular *streams of fire*, which, had they been permitted to last more than an instant, would certainly have fused the platinum points of the key, and *the helices became so hot* that the hand could not be kept on them. These effects could not have produced by the batteries. (Silliman, 1860, p. 97)
>
> *1. Observations at Christiania, Norway, by* Prof. CHRISTOPH HANSTEEN
>
> The effect of this aurora upon the telegraph lines in Norway was much greater than in France and Germany. The effect was noticed from the opening of the stations at 7 A.M. On the 29th communication was interrupted till 11 A.M. on almost all the lines; and likewise Sept. 2d, but with a long repetition after 2 P.M. Sept. 3d, only towards 8½ A.M. (Silliman, 1860, p. 388)
>
> *Effect on the Telegraph Wires, from the Comptes Rendus, T. XLIX, p. 365*
>
> From the evening of Aug. 28th until the morning of the 29th the needles of the magnetic telegraph at Paris were almost constantly in motion, as if a permanent current was passing through the telegraph wires. Business was therefore entirely interrupted, and could not be resumed until 11 A.M. Aug. 29th. The same effect was noticed on the telegraph lines from 4^h to 8^h on the morning of Sept. 2d, although no aurora was noticed on that day. Business was again interrupted, the needles were disturbed, and the bells were rung. (Silliman, 1860, p. 391)

Observations at Brussels by M. Quetelet

> About midnight Aug. 28th, the employees in the telegraph office at Brussels noticed signals from their bells, such as often occur during a storm. The employees in the offices at Mons, Antwerp, Gand and Ostend were also awakened by their bells, and enquired what was wanted. Communication with Paris, London, and Berlin were interrupted till $1^h 30^m$. Paris and London inquired of our operators if they saw a light in the heavens. The effect ceased at $1^h 30^m$ on all the lines except the submarine line from Ostend to Dover, which was charged with electricity throughout the entire morning. It was not till $3^h 30^m$, and after nearly doubling the battery, that communication was re-established.
>
> September 2, between 5^h and 6^h A.M., there was a second disturbance on all the telegraph lines, and communication between Brussels, Paris, and London was interrupted. (Silliman, 1860, p. 393)

1869

On May 30, 1869, during the aurora borealis which was visible from [in Switzerland] from 7 to 9 P.M., Heer observed that out of the sixteen lines which terminated in the telegraphic office in Basle, six were almost useless during the two hours that the phenomena lasted; on the others the telluric currents were not strong enough absolutely to interrupt communication (Angot, 1897, p. 141).

1872

Angot (1897, p. 142) reported the following disturbances to telegraph communication:

> The telluric currents attained an extraordinary development during the aurora of February 4, 1872, which we have mentioned as one of the most extensive known; it was seen in the whole of the West of Asia, in the North of Africa, throughout Europe, and on the Atlantic as far as Florida and Greenland; at the same time an aurora was observed in part of the southern hemisphere. The disturbances in telegraphic communication were not less extensive, and were observed with great care, in great part of Europe.... At the same time many of the submarine cables were so affected as to prevent the transmission of any messages; the disturbance was especially marked on the line from Lisbon to Gibraltar, on the Mediterranean cable, on the line from Suez to Aden, and from Aden to Bombay, and finally along the Transatlantic cable from Brest to Duxbury.

1881

Burbank (1905) reported that on 31 January the telegraph in England was affected by a current corresponding to a potential difference of 1.1 volts/km.

1882

According to W.H. Preece, Superintendent of the Telegraph, telluric currents observed in England on November 17 were five times as strong as the current usually employed in telegraphy. Communication was interrupted for as long as the disturbance lasted (Angot, 1897).

1891

Finn (1903, p. 23351) reported the following:

> During an electrical disturbance in 1891 similar to that through which we have just passed an electromotive force of 768 volts was recorded on the Western Union lines between New York and Buffalo, the circuits varying from 450 to 480 miles in length. On several occasions the strength of the earth current reached nearly 300 milliamperes. The average strength of the working currents does not exceed 35 milliamperes.

1892

A serious interruption of wire service occurred in the United States on 16 July (Sanders, 1961).

1894

Telegraph operators had been supplied with telephones and heard a variety of sounds produced by earth currents in the lines. Preece (1894, p. 554) recounted the following:

> Peculiar and weird sounds distinctly perceived, some highly-pitched musical notes, others resembling murmur of waves on a distant beach....The musical sounds would very much resemble those emitted by a number of sirens driven at first slowly, then increased until a 'screech' is produced, then again dying away. Duration of each averaged about twenty seconds.

1903

Lockyer (1903, p. 9) reported that on 31 October, practically the world's entire telegraph system was upset:

> The enormous development of the telegraph, telephone, cable, and other applications of electricity since the date of the last great magnetic storm has caused the disturbance to be more generally observed that was previously perhaps the case.
>
> Practically the worlds whole telegraph system was upset, and information from this country, France, and the United States and other lands shows that for several hours communication was almost completely interrupted.

1909

On 25 September, telegraph disturbances occurred in many parts of the world and were accompanied by auroral displays down to magnetic latitudes of 30°S (Silverman, 1995). In Western Australia, the telegraph lines between Perth and Kalgoorlie, a distance of 350 miles (560 km), and between Perth and Albany worked with all the batteries cut out for one half hour on the evening of 25 September. A magnificent aurora was also visible and the telegraph currents varied with the waxing and waning of the auroral light (Anonymous, 1909).

1921

Karsberg et al. (1959) reported that induced currents caused fires in telegraph equipment in Sweden. The exact date is not given, but Chapman and Bartels (1940) showed that a great magnetic storm occurred on 13–15 May 1921.

1938

On April 16, problems were encountered on the telegraph system in Norway (Harang, 1941).

1940

The magnetic storm and its effects are described in a series of papers by Davidson (1940), Germaine (1940), and McNish (1940). Germain (1940) described the effects of the storm on communications:

> When the magnetic storm of Sunday, March 24, 1940, developed, it immediately became apparent that it was unusual intensity and covered a wide area, interrupting nearly all overseas radiotelephone circuits, service to ships at sea, and a number of long distance land telephone and other communication services.
>
> While most of the serious effects on the services on the land lines took place between 9:00 a.m. and 3:00 p.m. Central Standard Time, the earth current disturbances persisted until the morning of the 25th and disturbances continued to intermittently affect overseas service for a number of days, with the Transatlantic service the most seriously affected.
>
> The most seriously affected sections of the country were on the Eastern Seaboard and in the states of North Dakota, Minnesota, Iowa, and Missouri. Less serous effects were felt as far south as Georgia and as far west as Arizona, Colorado, and Montana....
>
> During the storm, failures to less than 10 per cent of the long distance telephone land circuits throughout the country were experienced; however, out of Minneapolis about 80 per cent of the long distance circuits were out of service at one time during the day; while out of Chicago, more than 200 circuits could not be used between noon and about 2 p.m.

This storm also marks the first time that effects on power systems were reported. These were summarized by Davidson (1940, p. 365–366):

> The basic information was obtained from 22 power companies scattered about the United States and in Canada. Disturbances on power systems were reported by companies in New England, New York, eastern Pennsylvania and southern and eastern parts of Minnesota and in Ontario and Quebec. Reports of "No trouble" were received from systems in North Carolina, Georgia, Alabama, Illinois, Nebraska, Kansas, northern Minnesota, Utah, Idaho, Washington and California.
>
> Of the 10 systems on which disturbances were noted, the effects, listed in the inverse order of the number of cases reported, were:
>
> 1. Voltage dips ranging up to 10 per cent but generally of short duration (7 cases).
> 2. Tripping transformer banks by differential relay operation (5 cases involving 15 transformer banks).
> 3. Large increases or swings in reactive kva (4 cases).

1946

High neutral currents and tripping of transformers occurred in Ontario (Acres Consulting Services Ltd., 1975, p. 10):

> On March 28, 1946 between 0646 and 0717 hours EST, the Port Arthur transformer tripped six times and the Crow River transformer twice; it was reported that the telephone relays chattered quite vigorously just preceding each operation of the transformer differential relay.
>
> On September 22, 1946 a maximum current of 93 amp was measured in a transformer neutral at Port Arthur. The Port Arthur transformer differential relay protection operated eight times between 0515 and 0946 hours EST.

1957

On 21 January, disturbances occurred on power-feeding circuits on transatlantic submarine cables (Axe, 1968).

On 22 September, a power system 230kV breaker tripped because of saturation of transformer cores and excessive third harmonic currents in ground relays (Slothower and Albertson, 1967).

1958

On 10 February, a magnetic storm caused disruption of service on the TAT-1 transatlantic cable (Winckler et al., 1959). Abnormal power flows occurred in the power system in Minnesota (Slothower and Albertson, 1967). Lanzerotti and Gregori (1986) report that a power blackout occurred in the Toronto area. Acres Consulting Services Ltd. (1975, p. 10) reported the following:

> On February 10, 1958 at exactly 2104 EST, a transformer at Rayner Generating Station and the Port Arthur transformers T1 and T2 were simultaneously tripped at Port Arthur, loss of load occurred for about 4 minutes. However, Ontario Hydro does not know of any significant loss of revenue or any problems to the customers served by this station.

1960

On 13 November, disturbances occurred on power-feeding circuits on transatlantic cables (Axe, 1968). Thirty line circuit breakers were tripped on the power system in Sweden (Elovaara et al., 1992).

1972

A major magnetic storm on 4 August caused an outage of the L4 cable system in the continental United States (Anderson et al., 1974) and problems on power systems (Albertson and Thorson, 1974). Anderson et al. (1974, p. 1817) conducted a detailed analysis of the L4 outage and reported as follows:

> An outage of the Plano, Illinois, to Cascade, Iowa, link of the L4 coaxial cable occurred at about 2240 UT on 4 August 1972 during a large geomagnetic storm. The available geomagnetic data measured in North America, as well as data received from two satellite instruments, are analysed. These data show that, at the time of the L4 outage, the boundary of the magnetosphere was pushed to unusually low altitudes by a greatly enhanced solar wind. As a result, large,

> rapid changes of the earth's magnetic field changes at 2241 to 2242 UT were of such magnitude as to induce earth currents of sufficient strength to produce the L4 outage by causing a high-current shutdown of the system link. The geomagnetic disturbances that produced the shutdown were not of the auroral-electrojet type normally associated with disruptions of power systems.

Acres Consulting Services Ltd. (1975, p. B-1–B-3) summarized the effects on power systems in Canada, as follows:

> 1. Newfoundland and Labrador Power Commission, St. John's, Newfoundland
> The Corner Brook transformer tripped on differential relay (IAC, no harmonic restraint) and was returned to service several times. Long Harbour Terminal tripped as above twice.
>
> 2. Hydro-Québec, Montréal, Québec
> MW variations of a few per cent. Bersimis No. 2, on power-frequency control, had MW variation of 90 per cent. Significant variations of Mvars at all generating stations. Voltage variations of 3.6 per cent (735 kv sending end) to 5.7 per cent (315 kv receiving end). Shunt capacitor bank tripped off by overload protection at 1852 hours EDT on August 4.
>
> 3. Ontario Hydro, Toronto, Ontario
> Drop in power on the Ontario–Michigan interconnection from 540 to 480 Mw, and on the Ontario–New York State interconnection from 310 to 280 Mw. Voltage variations of up to 8 per cent. and a change in system frequency from 60.00 to 60.03 hertz.
>
> 4. Manitoba Hydro, Winnipeg, Manitoba
> Fluctuations on the U.S.A. Tie Line (La Verendrye) from 120 to 164 to 44 Mw and +25 to -100 Mvar. Drop in voltage at Grand Rapids, La Verendrye, Morden, Parkdale and Selkirk. At la Verendrye, SIC > 100 amp in neutral of 230/115-kv transformer and at Grand Rapids, SIC ~100 amp in transformer. System frequency varies from 60.00 to 59.95 to 60.08 hertz.
> Seven Sisters Unit No. 2 (25 Mw) tripped by generator field ground and overcurrent time A and C phase relays. Slave Falls Unit No.3 (9 Mw) tripped by undervoltage relay. Load Dispatch requested all generator units to maximum voltage boost at 1755 hours CDT, hoping to prevent loss of additional generator unit.

1980

In October, the 500 kV line from Manitoba to Minnesota was tripped (Aspnes et al., 1981).

1981

In April, the 500 kV line from Manitoba to Minnesota was tripped (Aspnes et al., 1981).

1982

On 13 July, 4 transformers and 15 lines were tripped in Sweden (Elovaara et al., 1992). On 14 July, railway signals were turned to red by the induced voltage (Wallerius, 1982)

1989

A great magnetic storm on 13 March 1989 caused a nine-hour blackout of the 21 000 MV Hydro-Québec electric power system. A vivid description of that failure is provided by Blais and Metsa (1993, p. 111) of Hydro-Québec:

> Telluric currents induced by the storm created harmonic voltages and currents of considerable intensity on the La Grande network. Voltage asymmetry on the 735-kV network reached 15%. Within less than a minute, the seven La Grande network static var compensators on line tripped one after the other....With loss of the last static var compensator, voltage dropped so drastically on the La Grande network (0.2 p.u.) that all five lines to Montréal tripped through loss of synchronism (virtual fault), and the entire network separated. The loss of 9450 MV of generation provoked a very rapid drop in frequency at load-centre substations. Automatic under-frequency load-shedding controls functioned properly, but they are not designed for recovery from a generation loss equivalent to about half system load. The rest of the grid collapsed piece by piece in 25 seconds.

Many other power utilities in North America experienced problems ranging from minor voltage fluctuations to tripping out of lines and capacitors. Boteler and Jansen van Beek (1993) provide a summary of where and when the effects occurre and the ionospheric current systems that caused them.

Bozoki et al. (1996) reported other power-system disturbances in North America in 1989. On 19 September, minor tank heating was observed on transformer T2 at Meadow Brook on the Allegheny Power System, and Public Service Electric & Gas Salem #2 generator step-up transformer was damaged; on 17–18 November, the Allegheny Power System Millvill SC was tripped by neutral unbalance protection, and a BC Hydro 138 kV line was tripped by overvoltage protection.

1991

Protective relay operations occurred on power systems in North America during geomagnetic disturbances on 22–24 March, 28 April, 16 May, 4–5, June, 10 June, and 28–29 October. The details are given by Bozoki et al. (1996, p. 735). Some of the effects observed include the following:

> Hydro-Quebec: Radisson dc harmonic overcurrent protection operated due to high 180 Hz component, resulting from high second harmonic content in 315 kV voltages. This event triggered loss of the Radisson-Sandy Pond Interconnection. Laurentides SVC tripped due to defective scheme.
>
> Allegheny Power System: Marlowe and Bedington SCs tripped by neutral unbalance relay. Numerous low voltage, zero-sequence capacitor neutral and transformer harmonic alarms. Both relays had notch filters. Guilford capacitor bank tripped by neutral unbalance relay; no notch filter. Gas accumulator relay tripped Meadowbrook T4; Meadow Brook T2 removed due to excessive noise and vibration.
>
> Central Hudson Gas & Electric: Hurley Ave. An induction disk overvoltage relay, used for neutral overcurrent protection, tripped a 115 kV capacitor bank.
>
> Virginia Power: Several SC trips were reported.

Southern California Edison: Fault recorder triggered at Serrano 230/500 kV station. Off-line analysis showed transformer neutral current of 150 to 200A.

Bonneville Power Administration: Five BPA capacitor bank trips — four by neutral unbalance and one by phase overcurrent relays.

BC Hydro: 138 kV line tripped on ground overcurrent.

APPENDIX B

Using the 'celestial power'

When people hear about natural electric fields, they commonly ask "Is there any way to use them as a power source?" The general answer is "No", because these electric fields occur too erratically.

However, at times in the past, the natural electric fields produced by geomagnetic storms were used to operate the telegraph system and send messages. Telegraph operators of the nineteenth century were as intrigued as we are at the idea of using this 'celestial power' and their descriptions make fascinating reading.

Prescott gave two accounts of the use of natural electric fields, one in a collection of reports on the great auroral disturbance of 1859 (Prescott, 1860) and the following, from his book on the history, theory, and practice of the electric telegraph (Prescott, 1866, p.320–321):

> We have seen from the foregoing examples that the aurora borealis produces remarkable effects upon the telegraph lines during its entire manifestation. We have, however, to record yet more wonderful effects of the aurora upon the wires; namely, *the use of the auroral current for transmitting and receiving telegraphic dispatches*. This almost incredible feat was accomplished in the forenoon of September 2 [1859], between the hours of half past eight and eleven o'clock, on the wires of the American Telegraph Company between Boston [Massachussetts] and Portland [Maine], upon the wires of the Old Colony and Fall River Railroad Company between South Braintree and Fall River [Massachussetts], and upon other lines in various parts of the country.
>
> The auroral influence was observed upon all the lines running out of the office in Boston at the hour of commencing business (eight o'clock, A.M.), and it continued so strong up to half past eight as to prevent any business being done; the ordinary current upon the wires being at times neutralized by the magnetism of the aurora, and at other times so greatly augmented as to render operations impracticable. At this juncture it was suggested that the batteries should be cut off, and the wires simply connected with the earth.
>
> It is proper to remark here, that, the current from the aurora coming in waves of greater or less intensity, there are times, both while the wave is approaching and while it is receding, when the instruments are enabled to work; but the time, varying according to the rapidity of the vibrations of the auroral bands, is only from a quarter of a minute to one minute in duration. Therefore, whatever business is done upon the wires during these displays has to be accomplished in brief intervals of from a quarter to half a minute in duration.
>
> During one of these intervals, the Boston operator said to the one at Portland: "Please cut off your battery, and let us see if we cannot work with the auroral current alone."
>
> The Portland operator replied: "I will do so. Will you do the same?"
>
> "I have already done so," was the answer. "We are working with the aid of the aurora alone. How do you receive my writing?"
>
> "Very well indeed," responds the operator at Portland; "much better than when the batteries were on; the current is steadier and more reliable. Suppose we continue to work so until the aurora subsides?"
>
> "Agreed," replied the Boston operator. "Are you ready for business?"
>
> "Yes; go ahead," was the answer.
>
> The Boston operator, Mr. Miliken, then commenced sending private despatches, which he was able to do much more satisfactorily than when the batteries were on, although, of course, not so well as he could have done with his own batteries without celestial assistance.

Nearly the same description is given in Prescott's earlier account (Prescott, 1860, p. 94–95), which continues as follows:

> While this singular phenomenon was taking place upon the wires between Boston and Portland, the operator at South Braintree — Miss Sarah B. Allen — informed me that she was working the wire between that station and Fall River — a distance of about forty miles south — with the auroral current alone. Since then I have visited Fall River and have the statement verified by the intelligent operator upon the railroad line at the dépôt in that village.
>
> The office at the dépôt is about a mile from the regular office in the village. The battery is kept at the latter place, but the operator at the dépôt is provided with a button or switch, by which he can throw the battery off the line, and put the wire in connection with the ground at pleasure. The battery at the other terminus of the line is at Boston, but the operator at South Braintree is furnished with a similar switch, which enables her to dispense with its use at pleasure. There are no intermediate batteries; consequently if Fall River operator puts his wire to the earth, and the South Braintree operator puts her wire to the earth, the line is without battery, and of course, without an electrical current.
>
> Such was the state of the line upon the 2d of September last, when for nearly two hours, they held communication over the wire with the aid of the celestial batteries alone!
>
> I have restricted myself in this article to facts observed by myself. I have stated nothing which I am not absolutely certain of, and which, if necessary, can be proved by a number of reliable witnesses.

Signals were also exchanged between Boston, Massachussetts, and Manchester, New Hampshire, and a message was sent from Philadelphia to Pittsburg by means of the earth current (Burbank, 1905).

Finn (1903, p. 23351) also reported the following:

> I notice a report made by Supt. Green, of the Northern Pacific Railroad service, in which he worked a line from St. Paul, Minn., to Dickinson, N. Dak., a distance of 650 miles, simply by the electricity taken from the atmosphere during the aurora borealis on Saturday, October 31 last. The strength of this current registered 23 millimeters positive back to the same figure negative.

Impact cratering on Earth

R.A.F. Grieve[1]

Grieve, R.A.F., 2001: Impact cratering on Earth; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 207–224.*

Abstract: Approximately 160 known terrestrial impact structures are distributed worldwide, with concentrations in geologically stable areas. The basic types are 'simple' craters, up to 4 km in diameter, with uplifted and overturned rimrocks surrounding a bowl-shaped depression, and 'complex' structures, generally over 3 km in diameter, with a central uplifted peak and/or ring, an annular trough, and a slumped rim. Morphology is used only to identify possible impact structures. The impact origin of smaller craters can often be determined by the presence of fragments of the impactor. At large craters, however, the impact commonly melts or completely vapourizes the impactor and verification is based on shock metamorphism. The formation of impact craters as small as 20 km reduces sunlight and disrupts temperature in a manner similar to a nuclear winter. A large impact event at the Cretaceous–Tertiary (K–T) boundary is the most likely cause of impactor mass extinctions that occurred at that time.

Résumé : Les quelque 160 astroblèmes terrestres recensés sur la planète se trouvent principalement dans des régions géologiquement stables. Les types de base sont des cratères «simples», faisant jusqu'à 4 km de diamètre, dont les bordures soulevées et renversées entourent une dépression en forme de cuvette, et les structures «complexes», dont le diamètre est généralement supérieur à 3 km, formées d'une aiguille ou d'un anneau en leur centre, d'un fossé annulaire et d'une bordure affaissée. La morphologie ne sert que pour mettre en évidence les éventuelles structures d'impact. La présence de fragments de météorite permet souvent de confirmer l'origine des petits cratères. Par contre, en ce qui a trait aux cratères de grand diamètre, l'impact du météorite a souvent fondu ou complètement pulvérisé l'objet sur lequel il est tombé et la vérification est alors basée sur le métamorphisme d'impact. La formation d'astroblèmes de diamètre aussi petit que 20 km réduit la luminosité du soleil et perturbe le climat de la même manière qu'un hiver nucléaire. L'impact d'un énorme météorite à la frontière du Crétacé et du Tertiaire est sans aucun doute à l'origine de l'extinction massive d'animaux survenue à cette époque.

---

[1] Continental Science Division, Geological Survey of Canada

INTRODUCTION

Not so long ago, impact phenomena were regarded as interesting, but certainly not important in the range of geological processes affecting Earth. Our concept of the importance of impact processes, however, has changed radically in recent years. This is due largely to planetary exploration, which has shown that nearly all planetary surfaces are heavily cratered from the impact of meteorites, asteroids, and comets. It is now evident from planetary bodies that have retained portions of their earliest surfaces that impact was a dominant geological process throughout the early solar system. For example, the oldest lunar surfaces are literally covered with impact craters, produced by an intense bombardment that lasted from 4.6 to approximately 3.9 billion years ago, and was at least 100 times higher than the present impact flux. Earth, as part of the solar system, underwent the same bombardment as the other planetary bodies, with impact cratering continuing throughout the last 3.9 billion years of solar-system history.

Impacts are now thought to have had a major effect upon the evolution of Earth. The best working hypothesis for the origin of the Earth's moon is that an object the size of Mars impacted with the primitive Earth. This resulted in vaporized material from the collision being placed into orbit around Earth; this material later condensed and fused to form the moon. The early, intense impact bombardment has also been proposed as an agent in the initiation of the compositional difference between the silica-rich continental crust and the iron-magnesium-rich oceanic crust. Heat generated by early impacts may have led to degassing and dehydration of Earth's primitive crust, thus contributing to the initial atmosphere and hydrosphere. Additionally, the impacting bodies themselves may have contributed to Earth's budget of atmospheric gases. This early bombardment would also have delayed the development and evolution of early life, with the largest impacts having the capacity to effectively sterilize the surface of the globe. Evidence has been found that at least one animal extinction, notably that of the dinosaurs and many other species 65 million years ago, is linked to global effects caused by a major impact event in more recent geological time. Of greater economic relevance, the vast copper-nickel deposits at Sudbury are possibly related to a large-scale impact that occurred 1850 million years ago. The faulting and fracturing associated with several impact structures in sedimentary rocks have also provided suitable traps for economic oil and gas deposits.

Most terrestrial impact craters have been destroyed by erosion, tectonics, and other geological processes. Some examples remain, preserved because of their young age, their large size, their occurrence in a relatively stable geological region, or their relatively rapid burial by younger protective sediments, which, in some cases, erosion has since removed. To date, approximately 160 impact craters have been identified on Earth. Almost all known craters have been recognized since 1950 and several new structures are found each year. Although these structures are distributed worldwide, they are concentrated in areas such as the Canadian Shield, whose relatively uneventful geological history over at least the past 500 million years has allowed preservation of some of the larger craters.

CRATER MORPHOLOGY

Extraterrestrial matter that has collided with Earth over more recent geological history ranges in size from meteoritic dust to bodies several kilometres in diameter. At the extremes, the effects upon Earth of such collisions range from negligible to catastrophic. The most obvious result of the larger collisions is seen in the spectrum of crater sizes and morphologies. Lunar studies have demonstrated that the morphology of impact craters changes systematically with crater diameter, which in turn depends largely on impactor size and velocity and on planetary gravity. This size-morphology relation can be seen in the fresh-looking lunar craters (Fig. 1). Relatively small-diameter craters are bowl shaped with raised rims ('simple' structure). As crater diameter increases, slumping of the inner walls and rebounding of the depressed floor create progressively larger rim terracing and central peaks ('complex' structure). At larger diameters, the single central peak is replaced by one or more peak rings, resulting in what is generally termed an 'impact basin'. This same progression in crater morphology is observed throughout the solar system, including on Earth, although terrestrial craters are less well preserved and, hence, more challenging to classify. One notable difference between lunar and terrestrial impact craters is the smaller diameter range for each morphological type on Earth. This difference is due to the different gravity fields of the two bodies. Comparative planetological studies indicate that transition diameters (from simple to complex to basin structures) are inversely related to planetary gravity. Thus, simple craters on the moon are larger than those on Earth because of the moon's weaker gravity.

On Earth, the basic impact crater types are 1) simple craters, up to 4 km in diameter, with uplifted and overturned rimrocks surrounding a bowl-shaped depression that is partly filled by broken and mixed rock (breccia), and 2) complex impact structures and basins, generally 3 km or more in diameter, with a distinct central uplift in the form of a peak and/or ring, a ring-shaped trough, and a slumped rim. The interiors of complex structures are partly filled with broken and mixed rock (breccia) and impact melt rocks.

CRATER IDENTIFICATION

Meteorite fragments recovered within or around a crater are the strongest evidence for an impact origin, but they cannot be obtained from every site. Fragments are found only at the smaller craters and they weather quickly in the terrestrial environment. For impact events on Earth that form craters larger than approximately 1 km across, the 'shock' pressures and temperatures produced upon impact are sufficient to completely melt and even vaporize the impacting body and some of the target rocks. For example, pressure from the impact of a chondritic (rich in silicate minerals) meteorite into granite at 25 km/s (a reasonable impact velocity for an asteroid impact on Earth) is 900 GPa or 9 000 000 bars. Thus, fragments of the impacting body are not found at these sites. In such cases, the recognition of a characteristic suite of rock and mineral deformations, termed 'shock metamorphism', that is produced uniquely by extreme shock pressures, is indicative of a hypervelocity impact origin. Examples of shock effects include conical fractures known as 'shatter cones', microscopic deformation features in minerals, particularly the development of so-called 'planar deformation' features in silicates, the occurrence of various solid-state glasses and high-pressure variations of minerals, and rocks melted by the intense heat of impact.

Some terrestrial structures have morphological characteristics consistent with simple or complex craters, but lack either pieces of the impacting body or definitive signs of shock metamorphism. This may be because suitable samples cannot be readily recovered as they are submerged beneath a deep, circular lake, buried under post-impact sedimentary rocks, or almost completely eroded. Continued investigation may yet produce evidence of shock metamorphism at some of these *possible* impact craters.

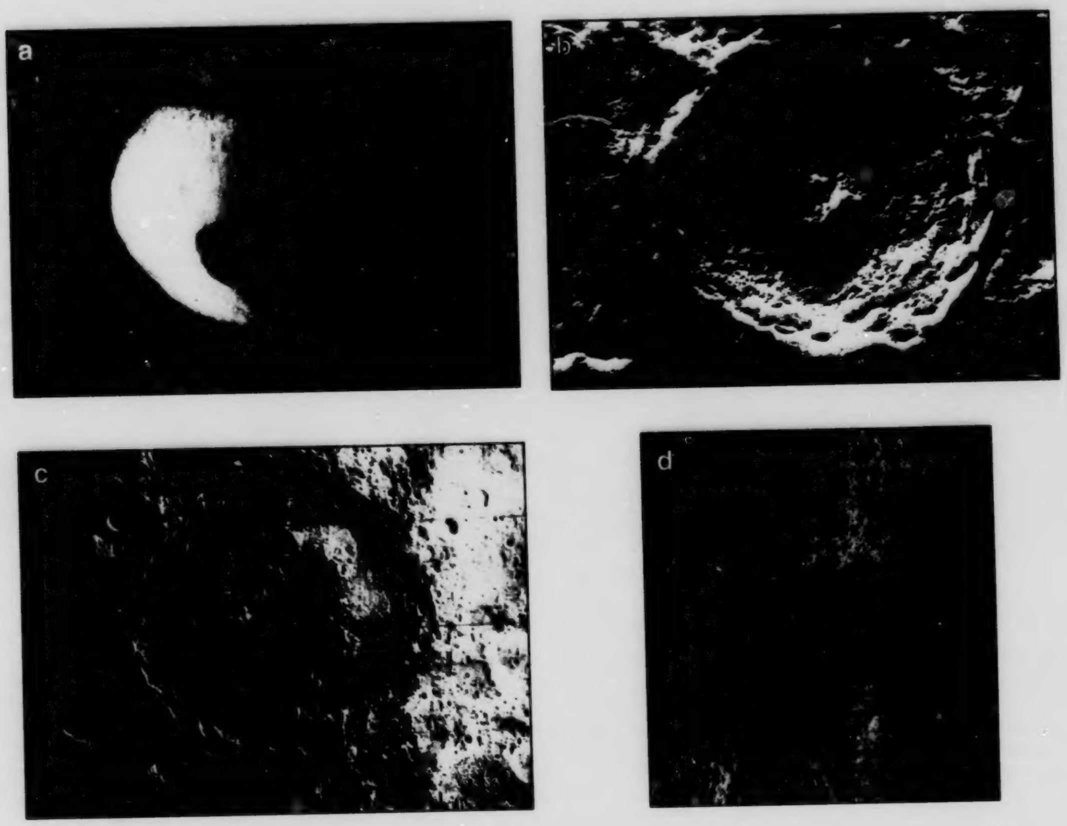


Figure 1. *Series of orbital photographs illustrating the progressive change of impact-crater morphology with increasing diameter on the moon.* ***a)*** *Alfrancus C, 10 km in diameter, is an example of a simple crater. Apollo 16 panoramic camera. NASA photograph 4615.* ***b)*** *Tycho, 85 km in diameter, is an example of a complex lunar crater. It has well developed wall terraces, a flat floor, and a fragmented central peak. Lunar Orbiter V, NASA photograph 125M.* ***c)*** *Schrödinger, 320 km in diameter, is an example of a lunar peak-ring basin. It has an almost complete interior peak ring, a flat floor, and terraced rim area. Lunar Orbiter IV, NASA photograph 9M.* ***d)*** *Orientale, approximately 900 km in diameter, is an example of a lunar multiring basin. It is the youngest and, thus, best preserved multiring basin on the moon. Lunar Orbiter IV, NASA photograph 194M.*

CRATER FORMATION

When extraterrestrial bodies tens of metres or larger impact with Earth, they are not slowed significantly by Earth's atmosphere. Because of their inherent high velocity, they have considerable kinetic energy. For example, the iron meteorite that formed the famous 1.2 km diameter Meteor or Barringer Crater in Arizona, in the United States (Fig. 2), was less than 50 m in diameter, but had a kinetic energy equivalent to that contained in approximately 50 megatons of TNT, equivalent to 3000 times the energy released in the Hiroshima atomic explosion. Upon impact, the bulk of this energy is transferred to the target rocks, leading to both the excavation of a crater and the production of diagnostic shock-metamorphic effects.

The transfer of energy to the target rocks occurs by radial spreading of a shock wave. The shock wave drives the target rocks radially downward and outward. A series of release waves follow the shock wave, returning the shocked target rocks to normal pressures. These release waves interact with the target rocks that are in motion, deflecting the movement of the rocks relatively close to the surface, so that their direction of movement is changed to upward and outward. This leads to the formation of a cavity, known as the 'transient cavity', by the combined forces of upward ejection and downward displacement of the target rocks. Portions of the walls of this cavity, which is about one third as deep as it is wide, are unstable and collapse inward, partly filling the crater with mixed and broken rock (breccia). Thus, a bowl-shaped or simple crater, partly filled with breccia, with a circular outline, an uplifted rim, and an exterior ejecta blanket, is produced.

The mechanisms of formation of complex craters and basins with uplifted central peaks and/or rings are more complicated. In these larger impacts, the initial stages of crater formation are similar to those of simple craters. As the resulting craters are so large, excavation is still proceeding at the outer edges of the transient cavity well after the cavity reaches its maximum depth. The downward-displaced floor of the cavity then rebounds, causing the formation of a central uplifted structure. The uplift of the floor also promotes collapse of the rim area, leading to a heavily modified, relatively shallow, complex structure with an uplifted center, a ring-like trough, and a collapsed and highly faulted rim area.

Although the number of known impact craters on Earth is relatively small, preserved craters are an extremely important resource for understanding impact phenomena. They provide the only accessible data on large-scale impacts currently available, and extensive geological, geophysical, and geochemical studies can be conducted at the sites. Earth's impact craters also provide important data on the structure of such landforms in all three dimensions. In some cases, the large size of terrestrial impact craters, up to approximately 300 km in diameter, requires satellite imagery and observation to provide a general view of their structure and the large-scale geological context.

IMPACT AS A HAZARD

The larger geological community has long discounted impact processes as a significant factor in Earth's more recent geological history. In 1980, this viewpoint was challenged severely when the high concentrations of siderophile elements (nickel, cobalt, iridium, platinum) present within Cretaceous–Tertiary (K–T) boundary sediments throughout the world were interpreted to have originated from a major impact event and this impact was suggested as the cause of mass animal extinctions that occurred at that time. Interestingly, the occurrence of the siderophile geochemical anomaly in K–T boundary sediments had been described some years previously. Similarly, the hypothesis that a major impact event could cause a mass extinction was not new. What was new was actually making the connection between the geochemical evidence at the K–T boundary and the impact process. This interpretation was consistent with our understanding of impact processes at the time and the known association of siderophile geochemical anomalies with some

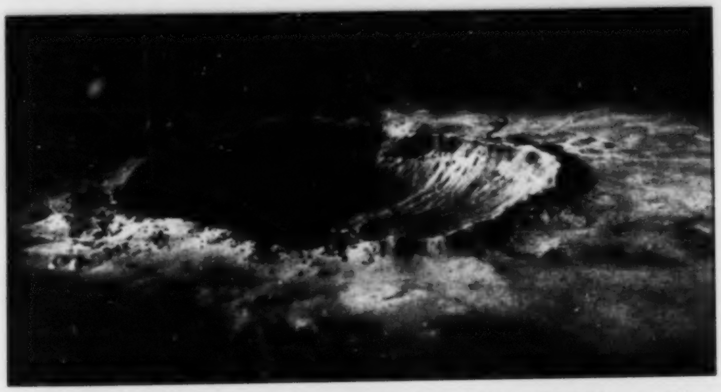


Figure 2. *Oblique aerial photograph of the 1.2 km, 50 000 year old Meteor or Barringer Crater in Arizona, United States, an example of a well preserved, terrestrial, simple crater. The hummocky terrain outside the rim is the remnant of the original ejecta blanket. Photograph courtesy of the United States Geological Survey*

impact craters, and was readily accepted by the impact and planetary geoscience community. The acceptance of the K–T impact hypothesis by the more general terrestrial geoscience community was less complete, and considerable controversy and debate were generated. Today, few scientists would deny the existence of abundant diagnostic evidence that a major impact event occurred at the K–T boundary. It is probably fair to say, however, that there is less consensus on the role of impact as a killing mechanism for the associated mass extinction event, with some scientists still having difficulty in accepting impact-related processes as the cause. Subsequent work has shown that the K–T boundary event was caused by the formation of the (now buried) 180 km diameter Chicxulub impact structure in the Yucatan Peninsula, Mexico.

Evidence of the K–T impact is recognizable globally, as large impact events have the capacity to produce an 'atmospheric blowout'. Essentially, a hole is blown out in the atmosphere above the impact site, permitting some impact materials to be dispersed globally not only by ballistic mechanisms, but also by entrainment in the impact fireball, which rises above the atmosphere. Unlike fine volcanic ash, these materials do not require atmospheric winds for dispersal and can encircle the globe in a relatively short time. They eventually settle upon the surface and leave a distinctive signature in the stratigraphic record. Model calculations indicate that a K–T-sized event is not necessary to produce atmospheric blowout. Relatively small impact events, resulting in impact structures in the 10 to 20 km size range, can produce atmospheric blowout. Currently, however, the K–T is the only stratigraphic boundary related directly to the fossil record that shows a clear link with a large-scale impact event. The involvement of impact in other boundary events in the terrestrial stratigraphic record has been suggested, but little definitive geochemical or physical evidence has been found.

From estimates of the terrestrial cratering rate in relatively recent geological time, the frequency of K–T-sized events on Earth is calculated to be in the order of one every 50 to 100 million years. Smaller, but still significant, impact events occur with shorter time scales (Fig. 3). Although such events may not be sufficient to produce mass extinctions, they will affect Earth's climate and biosphere to varying degrees. Model calculations suggest that dust thrown into the atmosphere from the formation of impact craters as small as 20 km in diameter could reduce sunlight and disrupt temperatures the way a nuclear winter would. Such impacts occur on Earth with a frequency of two or three every million years. The most recent known structure in this size range is Zhamanshin, Kazakhstan (part of the former USSR), with a diameter of 15 km and an age of about one million years (Fig. 160). Impacts of this scale are not likely to have a serious affect

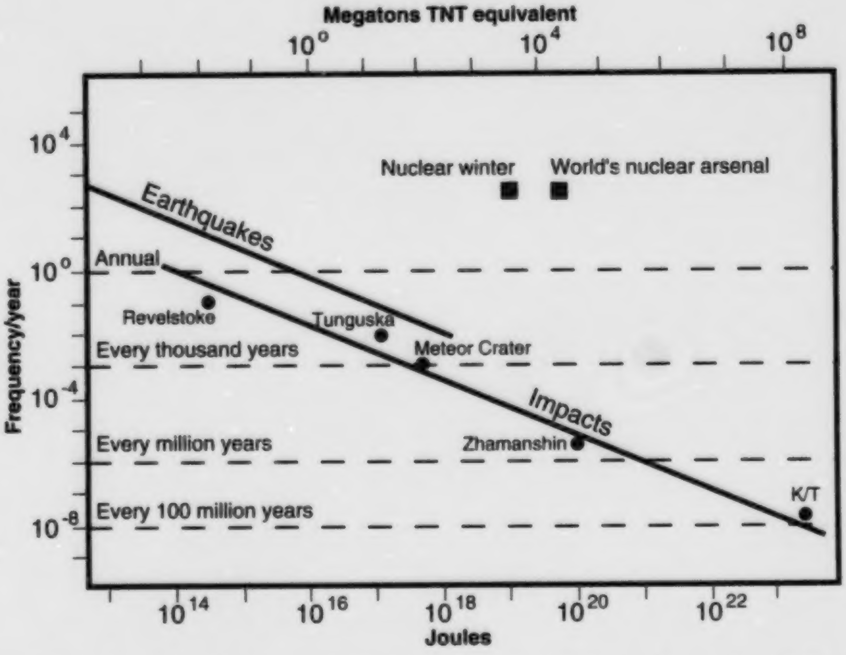


Figure 3. Frequency of impact against energy in Joules and megatons TNT equivalent. Some specific impact events are indicated with their relative frequency. Impacts at Revelstoke, British Columbia, and Tunguska, Russia, were atmospheric explosions. Also shown for comparison are the frequency-energy relationship for earthquakes and the energies associated with nuclear winter scenarios and the world's nuclear arsenal. An impact-induced nuclear winter is likely to occur on time scales of a few million years.

upon the biosphere and cause mass extinctions. The most fragile component of the present environment, however, is modern human civilization, which is highly dependent upon technology for its survival. Although we seldom think of civilization as spanning millions of years, there is little doubt that if our civilization survives long enough, it will suffer severely from or may even be destroyed by a major impact event.

In addition to throwing dust into the atmosphere, impacts will cause chemical changes in the atmosphere. These are related to the vaporization of the impacting body and a portion of the target rock. Considering only the contribution from the impacting body, recent calculations indicate that even relatively small impacting bodies, less than 0.5 km in diameter, that produce impact craters on the scale of 10 km in diameter, would inject five times more sulphur into the atmosphere than its present content. Larger impact events occurring on the time scale of a million years will inject enough sulphur to produce a drop in temperature of several degrees, causing a major change in climate. Additional effects on atmospheric chemistry include the potential for the destruction of the ozone layer from 'shock' heating of atmospheric nitrogen and the injection of fluorides from the vaporized impacting body. The threshold event for these effects would apparently occur on a time scale of hundreds of thousands of years. Additional fluorides can be produced from the target rocks, particularly if the impact occurs in a marine environment.

Marine impacts also have another consideration, namely the generation of tsunamis, which can be considerably larger than those generated by earthquakes. For example, an impact anywhere in the Atlantic Ocean by a body approximately 400 m in diameter would cause onshore wave runups of over 60 m, devastating the coasts on both sides of the ocean. The 1960 tsunami generated by a magnitude 8.6 earthquake in Chile is thought to have been the largest in the twentieth century. An impact-generated tsunami 10 times more powerful will occur with an average return time of 1000 years. Smaller impact-generated tsunamis will occur, but relatively infrequently, as there is a threshold size below which the impacting bodies burn up in the atmosphere and thus do not hit the Earth's surface. The threshold size at which atmospheric shielding is exceeded depends on the strength of the impacting body, which decreases from iron to stony meteorites to icy comets. For example, impacting iron bodies up to about 20 m in diameter will release their energy in the atmosphere and not reach the surface; on the other hand, comets as large as 200 m in diameter will release their energy in the atmosphere. Smaller impacting bodies can release their energy in the atmosphere as an air burst. The blast energy of a comet 200 m in diameter is on the order of several thousand megatons of TNT. Such air-burst explosions, fortunately, are not completely efficient at delivering their energy to the ground, because some of the initial explosive energy is directed into space.

Impact events can occur on human time scales. For example, the Tunguska, Siberia, event in 1908 was due to the atmospheric explosion, at an altitude of roughly 10 km, of a relatively small body approximately few tens of metres in diameter. From the energy required to produce the observed seismic disturbances, the energy released has been estimated to be equivalent to 10 to 100 megatons of TNT. Although the air blast devastated approximately 2000 km^2 of Siberian forest, there was no loss of human life because of the very sparse population. Events such as that at Tunguska occur on a time scale of hundreds of years, but the next time a Tunguska-sized event occurs, it could be over an urban centre. The rate at which the human population is increasing makes it all the more likely that the next Tunguska-sized event will cause loss of life.

Currently, some 1000 kg of interplanetary debris enter Earth's atmosphere daily. By many orders of magnitude, the bulk of this extraterrestrial matter is in the form of cosmic dust particles. Due to atmospheric effects and the high rate of geomorphological and geological processes on the Earth's surface, Earth retains a poor record of the impact flux. This is true even for larger objects that have the potential to form crater structures. For example, all known terrestrial craters with diameters less than 1 km were formed by iron or stony-iron meteorites, even though these classes of meteorites represent less than 50 per cent of known meteorite falls and finds. The higher chance of irons forming smaller craters is due to atmospheric crushing and breakup of weaker stony meteorites. Because of the effective shielding provided by the Earth's atmosphere and the resulting reduction in impact velocity of small meteorites by atmospheric resistance, the smallest interplanetary materials are of little hazard to individuals. However, buildings and individuals have been struck by small meteorites.

Although we can take some collective comfort in the fact that global, civilization-threatening impact events occur only every 500 000 years or so, it must be remembered that impact is a random process not only in space, but also in time. The next large impact with Earth could be an 'impact-winter'-producing event or even a K–T-sized event. To emphasize this point, in March 1989, an asteroid body named '1989 FC' passed within 700 000 km of Earth. This Earth-crossing body was not discovered until it had passed Earth. It is estimated to be in the 0.5 km size range, capable of producing a Zhamanshin-sized crater or a devastating tsunami. Although 700 000 km is a considerable distance, when orbital velocities are considered, it translates to a miss of the Earth by only a few hours.

MITIGATION STRATEGIES

No systems or procedures are currently in place specifically for reducing the effects of an impact. Routine minor impacts of small meteorites usually have a minimal effect. A truly major impact, such as that associated with the K–T event, is unmitigable, given the levels of energy involved. Mitigation of a moderate-sized impact may be possible, provided sufficient warning can be given. Forecasting such an event would be the responsibility of the astronomical community; the release of a formal warning and evacuation plan would fall to government emergency measures organizations.

Warning of an interplanetary collision is an astronomical and not a geological problem. It is also an international problem. At present, about 5% of the interplanetary objects with diameters greater than 1 km and orbits that intersect that of Earth have been observed. None of these will impact with Earth in the near future. A recent international Near-Earth-Object Detection Workshop recommended a 20-year international observing program, using six, 2 to 3 m telescopes equipped with special cameras and appropriate image- processing software. It was estimated that 95% of objects larger than 1 km in diameter would be discovered within the 20-year period. To be successful, any astronomical warning system will have to be worldwide. Similarly, the relatively few craters contained within the boundaries of a single nation and the wide variation in their state of preservation require that data from a larger number of craters and their lithological units need to be compiled to obtain a clearer understanding of the processes operating during and following large impact events. Canada, with its relatively high number of craters, has an important role in this area of research.

SUGGESTED READING

Chapman C.R. and Morrison, D.
1989: Cosmic Catastrophes; Plenum Press, New York, 302 p.
Gehrels, T. (ed.)
1994: Hazards Due to Comets and Asteroids; University of Arizona Press, Tucson, Arizona, 1300 p.
Grieve, R.A.F.
1990: Impact cratering on the Earth; Scientific American, v. 262, p. 66–73.
Hildebrand, A.R.
1993: The Cretaceous/Tertiary boundary impact (or the dinosaurs didn't have a chance); Journal of the Royal Astronomical Society of Canada, v. 87, p. 77–118.

APPENDIX A

The impact structures of Canada

General summaries of the known impact structures in Canada are listed below. The locations of the structure are shown in Figure A-1.

Brent

The Brent impact structure (lat. 46°05′N; long. 78°29′W) is located near the northern boundary of Algonquin Provincial Park, Ontario (Fig. A-1). Named after the village of Brent, 4 km south, the structure was first noticed in 1951 as an anomalous circular feature on aerial photographs.

Brent is visible as a circular depression about 3 km in diameter and 60 m deep, with the terrain inside the depression being notably smoother than the surrounding area (Fig. A-2). The structure is almost completely filled in by post-impact sedimentary rocks. The original rim has been completely removed by erosion and the original rim diameter is estimated to have been approximately 3.8 km. Information on the subsurface geological character of Brent comes from the about 5 km of core recovered from 12 holes drilled between 1955 and 1967. The central holes penetrated glacial drift, up to 263 m of undeformed, post-impact sedimentary rocks and a breccia lens with a maximum thickness of 630 m, before passing into fractured and shocked target rocks. On the basis of geochemistry, it has been suggested that the impacting body was a chondrite (rich in silicate minerals). The age of Brent is listed at 450 ± 30 Ma, from isotopic analyses of the impact melt rocks.

Carswell

The Carswell impact structure (lat. 53°27′N; long. 109°30′W) is in northern Saskatchewan, approximately 120 km south of Uranium City (Fig. A-1). It is named after Carswell Lake, which transects its the northeastern boundary. Breccias, shatter cones, and other shock-metamorphic features have all been described at the site.

Carswell is a heavily eroded, complex impact structure. It is barely visible as a vague circular feature, corresponding to the outcrop of ridges of dolomite of the Carswell Formation (Fig. A-3). This ring is about 39 km in outer diameter and is generally given as the diameter of the structure. Essentially no crater-filling materials remain, beyond dyke-like and other bodies of so-called 'Cluff Breccia'. From this, it is concluded that the Carswell impact structure has been eroded to the floor of the original crater. Isotopic analyses of two impact melt samples gives ages of 115 ± 10 Ma.

Charlevoix

The Charlevoix impact structure (lat. 47°32′N; long. 70°18′W) is on the north shore of the St. Lawrence River in southern Quebec, 105 km north east of Québec (Fig. A-1). It is named after the county in which it is located. Shatter cones were discovered during regional mapping of the area.

Charlevoix is obvious as a semicircular area on the north shore of the St. Lawrence River that differs markedly from the topography of the region (Fig. A-4). Its eastwardly extension to the south shore is not visible. It is a heavily eroded structure, with essentially all crater-filling products having been removed. Its age is generally listed as 357 ± 15 Ma, on the basis of isotopic data from impact melt rock.

Couture

The Couture impact structure (lat. 60°8′N; long. 75°20′W) is on the Ungava Peninsula of northern Quebec, approximately 150 km southwest of the Nouveau Québec impact structure and 100 km east of Povungnituk (Fig. A-1). The impact structure is contained wholly within Couture Lake. An impact origin was first suggested in 1955 from the anomalous round shape of the lake as seen on a map. In 1963, samples of breccia were discovered.

Couture is roughly circular with a diameter of about 13 km (Fig. A-5). An island-free central area approximately 10 km in diameter is surrounded by a ring of islands and peninsulas. The impact structure is about 8 km in diameter. Breccia boulders up to 1to 5 m in diameter occur on some islands and in glacial deposits on the western shore. They also occur in a fan extending over 10 km to the west and are presumed to have been scoured from the lake bottom by glacial action. The age of the Couture impact structure has been determined by isotopic analyses of three samples of impact melt rocks to be 425 ± 25 Ma.

Deep Bay

The Deep Bay impact structure (lat. 56°24′N; long. 103°00′W) is at the southern end of Reindeer Lake in northeastern Saskatchewan, 100 km northwest of Flin Flon, Manitoba (Fig. A-1). It takes its name from the unusually

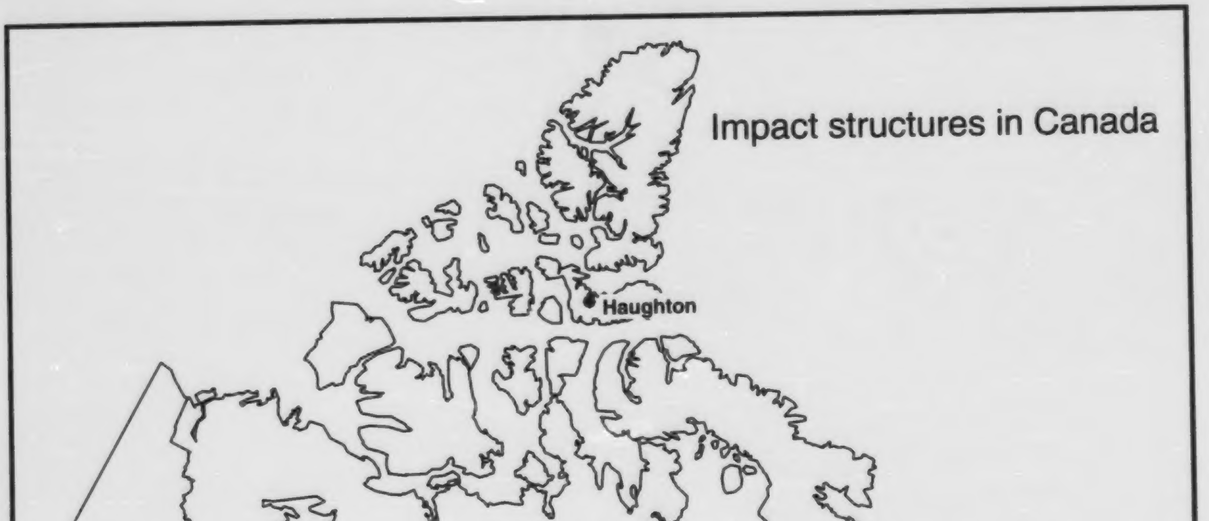

Impact structures in Canada
Haughton
Nicholson
Pilot
Steen River
Carswell
Gow
Deep Bay
Nouveau Québec
Couture
La Moinerie
Mistastin
Eau Claire (west)
Eau Claire (east)
Manicouagan
île Rouleau
Presqu'île
Charlevoix
Montagnais
Elbow
Eagle Butte
Maple Creek
Saint Martin
Viewfield
West Hawk
Slate Islands
Sudbury
Wanapitei
Brent
Holleford

Figure A-1. Impact structures in Canada.

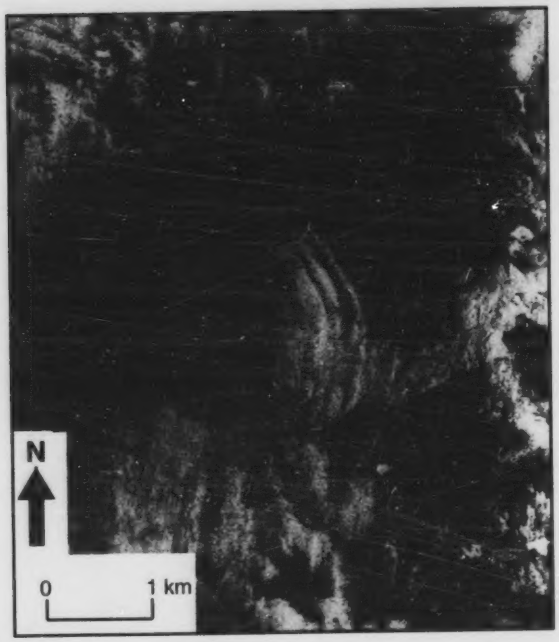


Figure A-2. Brent impact structure. National Airphoto Library photograph

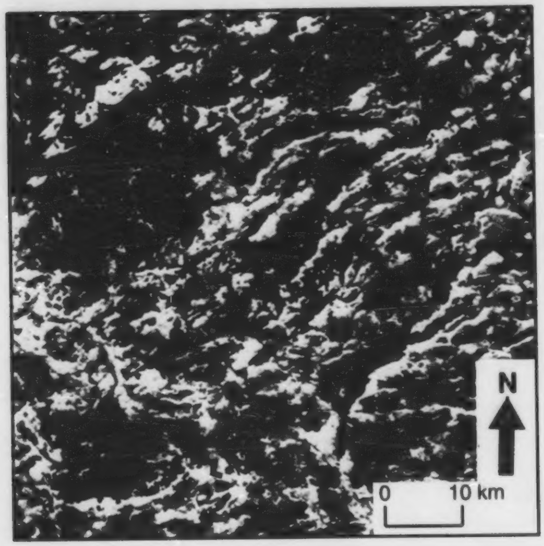


Figure A-3. Carswell impact structure. Landsat photograph courtesy of the National Aeronautics and Space Administration

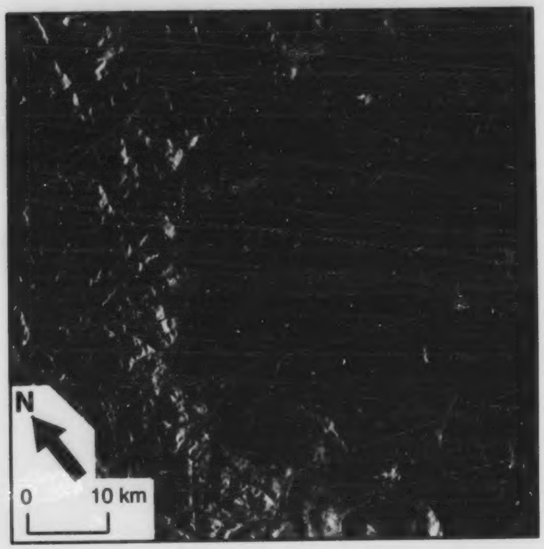


Figure A-4. Charlevoix impact structure. Courtesy of the National Aeronautics and Space Administration

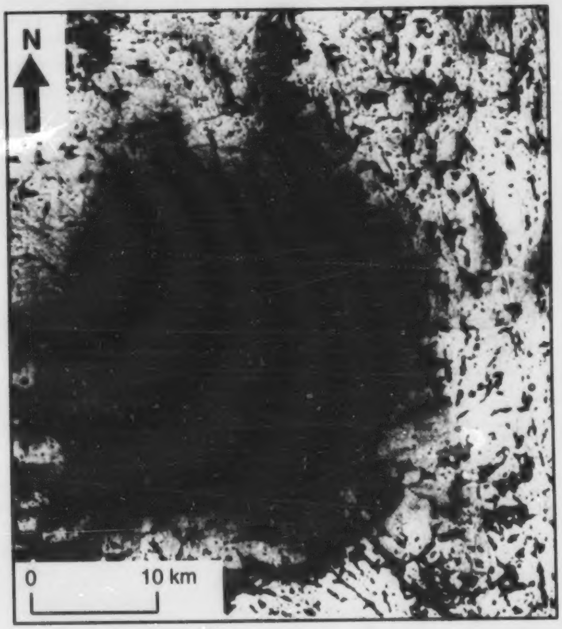


Figure A-5. Couture impact structure. National Airphoto Library photograph deep water in the bay, which marks the site of the structure and has a maximum depth of 220 m, compared to a maximum depth of about 45 m for the remainder of the lake. It was first suggested as an impact crater 1951, and detailed geological, geophysical, and drilling investigations were carried out at the site between 1956 and 1963.

The unusual circular shape and island-free character of Deep Bay compared to other bodies of water in the area are clearly evident (Fig. A-6). The circular shoreline has a diameter of about 11 km and is partly surrounded by a discontinuous ridge that locally reaches heights of 100 m above the lake level . This ridge is best expressed in the east and southeast and has a diameter of about 13 km, which is believed to be the original diameter of the impact structure. On the basis of an estimated paleontological age (using fossils) for the post-impact sedimentary rocks, the age of the structure is estimated at 100 ± 50 Ma.

Eagle Butte

The Eagle Butte impact structure (lat. 49°42′N; long. 110°30′W) is in southern Alberta, 50 km east of Medicine Hat (Fig. A-1). It takes its name from a nearby community. The structural anomaly at Eagle Butte has been known since the 1920s and past explanations for its origin have been varied. It was only in 1985 that evidence of shock-metamorphic effects, in the form of shatter cones and planar fractures in quartz, was discovered.

The structure is covered by glacial overburden and has a very subtle surface expression. The area is portrayed on geological maps as circular to oval, about 16 km in diameter, with complex faulting. The outer diameter of the structure, as defined by the structurally disturbed area at the surface and by seismic reflection data, is taken to be 19 km. The subsurface character is that of a complex structure, with a structural rim, ring-like trough, and central uplift. Considerable geological and, in particular, geophysical data have been acquired over the structure, as producing oil wells are found outside the structure. The exact age of the Eagle Butte structure is unknown, but is generally given as younger than 65 Ma.

Eau Claire

Eau Claire consists of two roughly circular lobes separated by a barrier of islands (Fig. A-1, A-7). It is the result of two impact structures, Eau Claire East (lat. 56°05′N; long. 74°07′W) and Eau Claire West (lat. 56°13′N; long. 74°30′W). The Eau Claire structures are located 200 km north of Poste-de-la-Baleine (Kuujjuaraapik) and 110 km east of Hudson Bay, in northern Quebec. The paired structures at the site are unique in North America. In 1958, geologists working on the ring of islands in the western lake observed breccia that was later recognized as impact melt rocks.

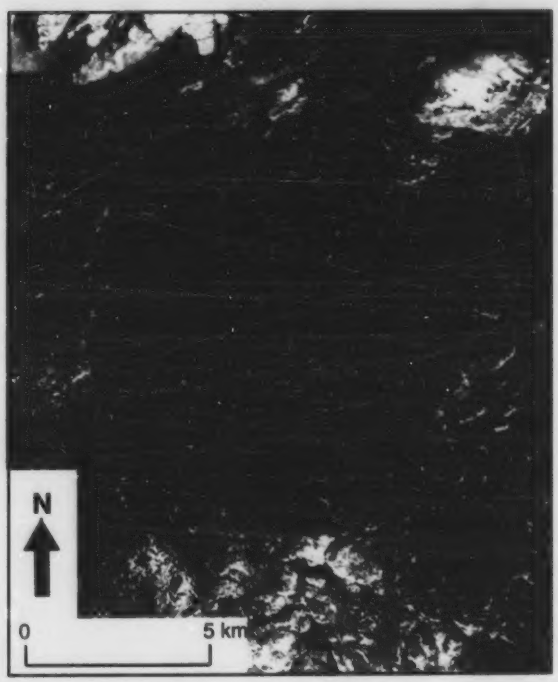


Figure A-6. Deep Bay impact structure. National Airphoto Library photograph

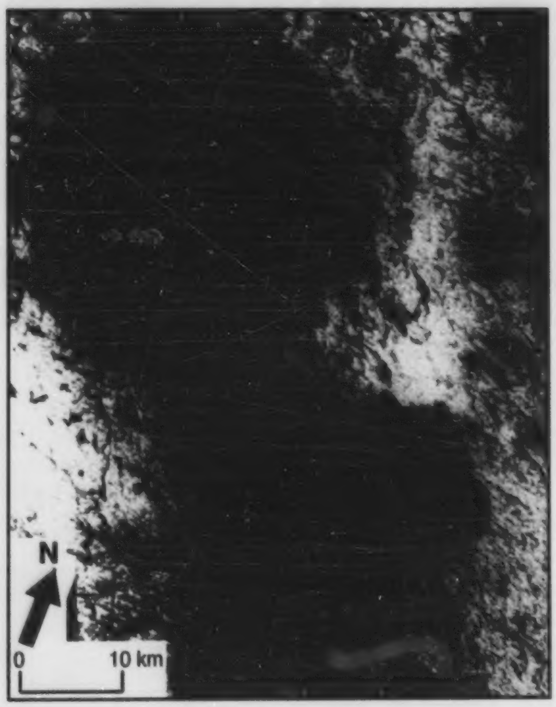


Figure A-7. Eau Claire impact structures. Shuttle image courtesy of the National Aeronautics and Space Administration

The western lake is about 25 km in diameter and its shoreline has numerous small islands. The land rises to a poorly defined and interrupted rim 5 to 10 km from the shore. The overall diameter of the impact structure is about 32 km. A prominent ring of islands approximately 16 km in diameter is found in the western lake. In the centre, a few small islands are less than 100 m in diameter. The eastern lake is about 24 km in diameter, with an island-studded shoreline. The island-free portion of the lake represents the 22 km diameter impact structure, which has no surface expression. Drillhole data indicate the presence of a central uplift beneath the crater-filling material.

Rocks from the two structures yield similar isotopic ages, suggesting that both structures are the result of simultaneous impacts, with the age of impact given generally as 290 ± 20 Ma. The size and separation of the two structures suggest that the impacting bodies were a binary pair.

Elbow

The Elbow impact structure (lat. 50°59′N; long. 106°43′W), located near a bend in Lake Diefenbaker in south-central Saskatchewan, is named after the small town of Elbow (Fig. A-1). It was first noticed as a structural anomaly and was initially drilled for hydrocarbons by Imperial Oil in 1945.

The Elbow structure has a slight surface expression as a flat-topped rise some 30 m above the surrounding land. Additional wells were drilled in 1953 and 1955, but all were unsuccessful at delineating commercial hydrocarbon production. The origin of the structure at Elbow was ascribed initially to folding or salt tectonism. The structure is largely delineated on the basis of seismic imaging and gravity data. A central structural uplift 3 km in diameter is surrounded by an annular low. The overall diameter is 8 km. Early reports of material recovered from the wells noted 'frosty' sand grains. More recent work in the 1990s indicates that these sand grains are actually the mineral quartz and the 'frosting' is, in fact, due to the occurrence of shock-metamorphic microscopic planar deformation features. The occurrence of shock features and the form of the Elbow structure indicate that it is a complex impact structure. Its age is poorly known, but on the basis of missing strata, the impact event is estimated to have occurred approximately 395 ± 25 Ma ago.

Gow

The Gow impact structure (lat. 56°27′N; long. 104°29′W) is located in northeastern Saskatchewan about 90 km west of the Deep Bay structure (Fig. A-1). Gow Lake occupies the structure and has an anomalous circular shape that is unique to an area where lakes are generally linear.

Gow Lake itself is roughly circular, being elongated to the northeast due to glaciation (Fig. A-8). It has a diameter of about 4 km and contains an elliptical central island, 1 km by 2 km, which represents an eroded central peak. The surrounding terrain reaches about 25 m above lake level at 0.5 to 1 km beyond the shoreline. This is taken to represent the eroded rim, giving the Gow Lake structure an estimated original diameter of about 5 km. Thus, Gow is one of the smallest known complex impact structures in crystalline target rocks. The isotopic analysis of one sample of the impact melt rock suggests an age of approximately 250 Ma.

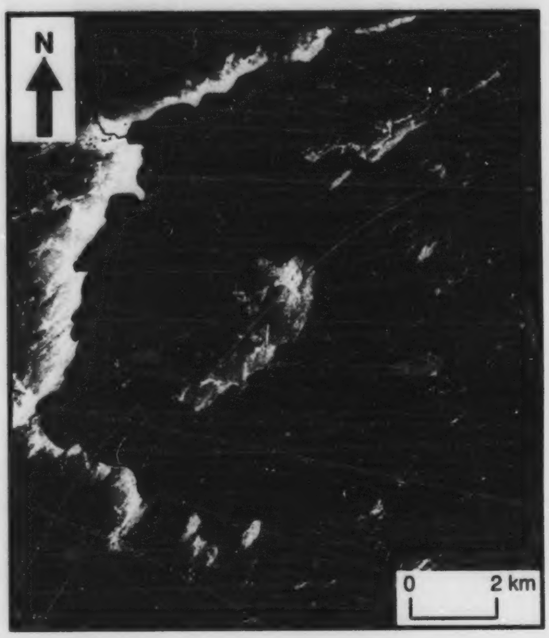


Figure A-8. *Gow impact structure. National Airphoto Library photograph*

Haughton

The Haughton impact structure (lat. 75°22′N; long. 89°40′W) is in the Canadian Arctic, on the west side of Devon Island (Fig. A-1). Shatter cones were discovered during geological exploration in 1974. More extensive mapping resulted in the discovery of extensive breccia partly filling the centre of the structure as well as other evidence of shock metamorphism.

Haughton is a circular area 16 to 17 km in diameter (Fig. A-9). Additional mapping and structural analysis revised the estimated diameter upwards to 20.5 km. Breccia deposits cover about 56 km² in the centre of the structure. Their maximum thickness is estimated to be at least 90 m and perhaps as much as 125 m. Isotopic analysis gives an age of 23.4 ± 1 Ma for the structure.

Holleford

The Holleford impact structure (lat. 44°28′N; long. 76°38′W) is 27 km northwest of Kingston, in southern Ontario (Fig. A-1). It takes its name from the community of Holleford, which lies partly within the structure. Holleford was first noted in 1955, during a systematic search of large-scale aerial photographs of the Canadian Shield for circular features of possible impact origin. Three holes were drilled into the Holleford structure in 1956–1957 as part of a combined geological and geophysical investigation of the structure.

The Holleford structure appears as a circular depression 2.3 km in diameter and about 30 m deep (Fig. A-10). The estimated diameter is inferred from the change in dip of post-impact limestone, which has partly filled the structure. The age is not known exactly, but is generally given as 550 ± 100 Ma, a minimum age based upon post-impact material filling the crater.

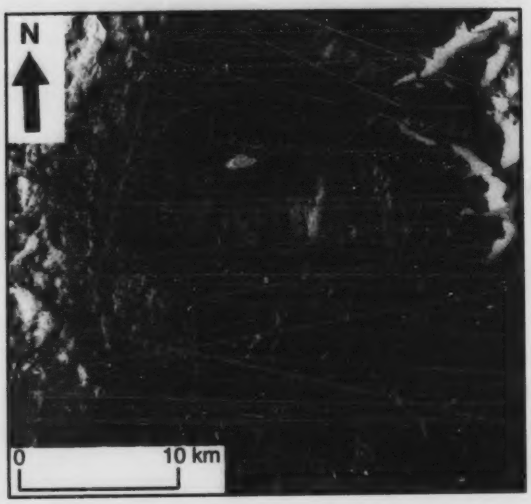


Figure A-9. Haughton impact structure (C-band radar image courtesy of the Earth Physics Branch)

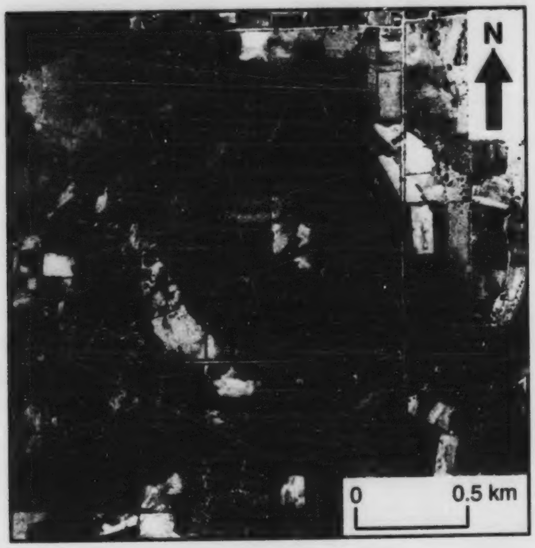


Figure A-10. Holleford impact structure. National Airphoto Library photograph

Île Rouleau

The Île Rouleau impact structure (lat. 50°41′N; long. 73°53′W) is in Lake Mistassini in central Quebec, approximately 90 km north of Chibougamau (Fig. A-1). It was discovered during regional mapping in 1973 and is named after the island that is its the sole known expression.

The island has a maximum diameter of 1 km and is bounded by 7 to 15 m high cliffs on its north and west shores (Fig. A-11). No indication of lithological effects of the impact event have been recognized on the larger Manitounouc Island to the east. It is assumed that Rouleau Island represents the eroded remnant of a central peak. From its dimensions and the general lack of obvious effects on Manitounouc Island, it is assumed that the Île Rouleau structure is about 4 km in diameter. The rocks on Rouleau Island are brecciated, faulted, and folded. Breccia dykes occur throughout the island and are up to 6 m thick. Shatter cones are well developed on the eastern and southwestern shores of the island. The age of the impact structure is unknown, but is given at less than 300 Ma, which must be considered a best guess.

La Moinerie

The La Moinerie impact structure (lat. 57°26′N; long. 66°37′W) is located in northeastern Quebec, about 125 km southeast of Kuujjak (Fig. A-1). It takes its name from the lake that occupies the structure. The unusual, quasi-circular to square form of the lake attracted attention as a possible impact site, and impact breccia rocks were discovered on the northern shoreline.

It is assumed that the impact structure lies totally within the lake, a situation analogous to that of Couture (Fig. A-12). No obvious rim is know and no other morphometric data exist

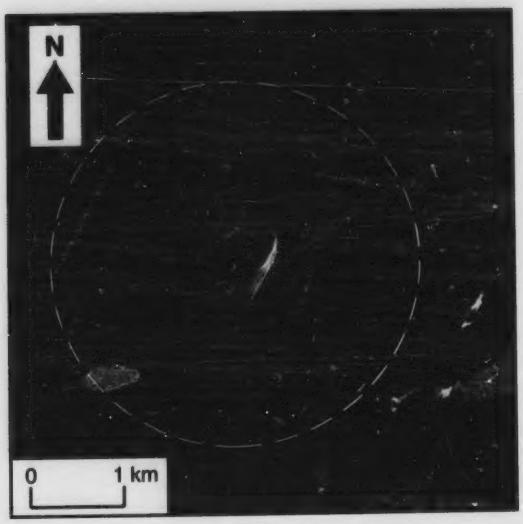


Figure A-11. Île Rouleau impact structure. National Airphoto Library photograph

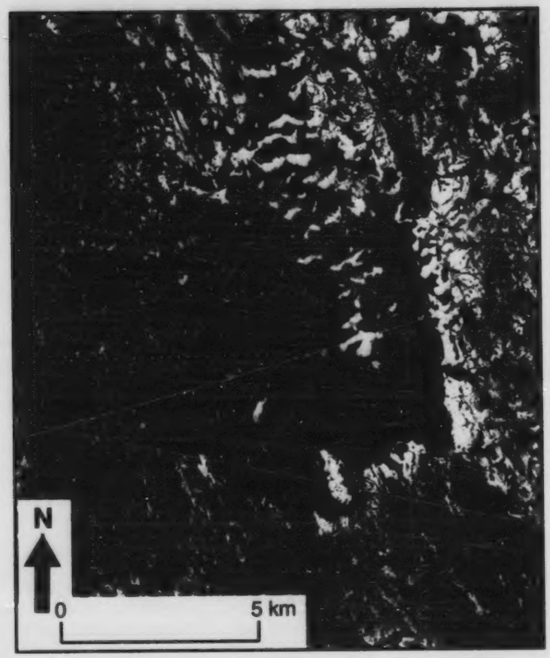


Figure A-12. La Moinerie impact structure. National Airphoto Library photograph

beyond the presence of a number of small central islands. The structure is generally assumed to be a complex impact structure, about 8 km in diameter, with a central uplift. Impact rock types are limited to glacial drift from the north side of the lake. Isotopic dating of three impact melt rocks indicate that the structure is 400 ± 20 Ma.

Manicouagan

The Manicouagan impact structure is in central Quebec, 180 km northwest of Sept-Îles (Fig. A-1). The unusual circular shape of the ring-shaped trough now occupied by the Manicouagan hydro-electric reservoir led to the suggestion in 1956 that it was a possible impact crater (Fig. A-13). A geophysical survey was carried out in 1963 and led to the discovery of shock metamorphism within rocks of the central area of the structure.

The structure is very obvious because of the Manicouagan reservoir and is a common photographic subject for space-shuttle astronauts. The Manicouagan reservoir is about 65 km in outer diameter and 5 to 10 km wide. In early studies, the outer diameter was given as the size of the impact structure. Recent studies, however, have established that the reservoir lies within the limits of the original impact structure. The original diameter is estimated to be 100 km, on the basis of topographic, geological, and geophysical evidence. Impact melt rocks are confined to the inner plateau, where they form a sheet up to 230 m thick overlying a basal breccia layer or basement rocks. Isotopic dating of the impact melt rocks provides an age of 214 ± 1 Ma.

Maple Creek

The Maple Creek impact structure (lat. 49°48′N; long. 109°06′W) is located in the northern flank of the Cypress Hills of southwestern Saskatchewan (Fig. A-1). It is named after the nearest town, Maple Creek, which is located some 30 km west-northwest of the impact structure.

It was first noticed in the mid-1940s as an area with high, deformed rocks from some depth that are now (barely) exposed at the surface. Hydrocarbon drilling defined a structurally complex central area of uplifted and missing strata and seismic data indicated a structural disturbance with an overall diameter of 6 km. No commercial hydrocarbons were encountered. All deformation is confined to the upper 1 km of the strata in the area. There is a central complex structural uplift 3 km in diameter, where geological modelling indicates that the rocks have been uplifted approximately 700 m. This uplift is surrounded by an annular trough. In other words, Maple Creek has the form of a complex impact structure. Recently, shock-metamorphic planar deformation features have been identified in rare grains of the mineral quartz in strata from the central uplift. The age of the impact event is not well constrained, but on the basis of the ages of the strata affected by the event, it is estimated to be less than 75 Ma.

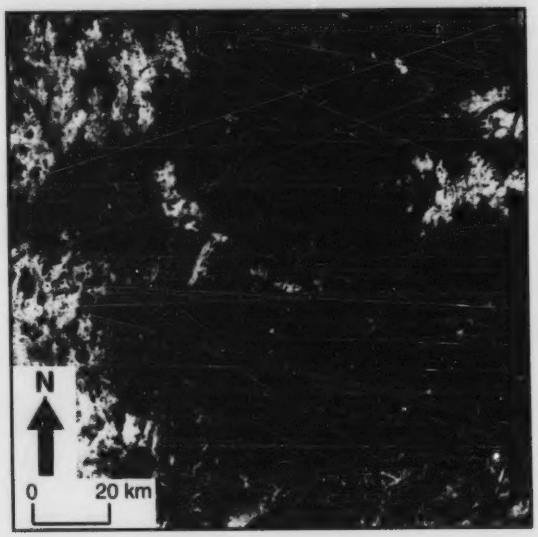


Figure A-13. Manicouagan impact structure. Landsat image courtesy of the National Aeronautics and Space Administration

Mistastin

The Mistastin impact structure (lat. 55°53′N; long. 63°18′W) is in central Labrador, 120 km north of Schefferville, Quebec (Fig. A-1). It contains Lake Mistastin. Abundant evidence of shock metamorphism was discovered in the late 1960s and the site was assigned an impact origin.

Lake Mistastin's oval shape, drawn out to the northeast, is anomalous in the area (Fig. A-14). The Lake is contained within a general depression surrounded by a low ring of hills about 28 km in diameter. It contains a horseshoe-shaped island about 3 km in diameter, which rises approximately 130 m above lake level. The island consists largely of shocked basement rocks and is the eroded remnant of a central uplift. The Mistastin impact structure is eroded to close to the original floor and has the form of a complex central-uplift structure. Isolated outcrops of impact melt rock occur around the lakeshore and the central island, with a major outcrop, 'Discovery Hill' in the southwest exposing a vertical section of approximately 80 m of impact melt rocks. It has been suggested that the impacting body was an iron. On the basis of the isotopic analyses of impact melt rocks, the Mistastin impact structure is dated at 38 ± 4 Ma.

Montagnais

The Montagnais impact structure (lat. 42°53′N; long. 64°13′W) is located on the continental shelf off Nova Scotia, 200 km south of Halifax (Fig. A-1). It is named after an exploratory oil well, Montagnais 1-94, drilled by the Union Oil Company, which intersected the structure.

Montagnais 1-94 was drilled into the structure in 1974 and encountered various units of conglomerate, breccia, and horizons of basaltic material. These were reinterpreted in terms of impact products following the discovery of shock-metamorphic features in samples from the well cuttings. Water depths in the area are about 115 m. Details of its morphology are not well known as it is submerged, and knowledge about the character of its morphological elements is largely based on geophysical data. Seismic profiles indicate a structure at least 45 km in diameter, with a central uplifted area about 11.5 km in diameter at its upper surface. A ring-like trough around the central uplift is filled with material, interpreted as breccia. Montagnais 1-94, which was drilled near the centre of the central uplift, penetrated 510 m of post-impact sedimentary rocks, and 552 m of various breccia units, before penetrating uplifted and deformed Meguma basement rock. Isotopic analyses of seven samples of impact melt rock give an estimated age of 50.5 ± 0.8 Ma.

Nicholson

The Nicholson impact structure (lat. 62°40′N; long. 102°41′W) is in the southern Northwest Territories, 320 km east of Fort Reliance (Fig. A-1). It is largely expressed by Nicholson Lake. Evidence of shock metamorphism at the site was discovered in 1965.

Nicholson Lake has roughly the shape of a rounded diamond, with a number of central islands and a large promontory in the west (Fig. A-15). The form of the structure is assumed to be that of a complex central-uplift structure. It is clearly heavily eroded and its original diameter is not well constrained, but is generally given as 12.5 km, very slightly

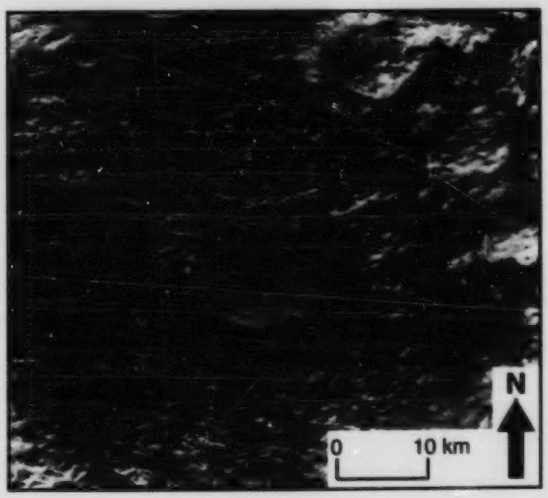


Figure A-14. Mistastin impact structure. Landsat image courtesy of the National Aeronautics and Space Administration

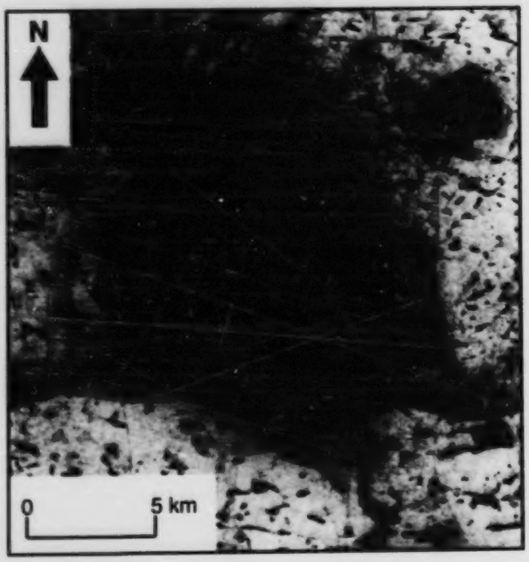


Figure A-15. Nicholson impact structure. National Airphoto Library photograph larger than the diameter of the lake. Outcrops of dolomite and limestone occur on the western and southwestern sides of the lake and represent remnants of the cover rocks downfaulted to their present position in the area of the ring-like trough of the impact structure. A single isotopic analysis of an impact melt rock gives a maximum age for the impact event of less than 400 Ma.

Nouveau Québec

The Nouveau Québec impact structure (lat. 61°16′N; long. 73°40′W) is on the Ungava Peninsula, northern Quebec, about 150 km southeast of Salluit (Fig. A-1). It was originally named 'Chubb crater', but its name was changed in 1954. Although the Nouveau Québec structure was the first in Canada proposed to be of impact origin, it was only in the early 1960s that the first shock-metamorphosed sample was recovered.

The nearly circular Nouveau Québec structure contrasts sharply with the linear lakes of the general area (Fig. A-16). The partly infilled, bowl-shaped lake is 2.8 km in diameter and 250 m deep. The rim is 3.4 km in diameter and rises to a maximum of 163 m above the lake level and 80 to 100 m above the regional terrain outside the structure. Isotopic analyses of three impact melt rocks date the structure at 1.4 ± 0.1 Ma.

Pilot

The Pilot impact structure (lat. 60°17′N; long. 111°00′W) is about 60 km northeast of Fort Smith, Northwest Territories (Fig. A-1). It takes its name from Pilot Lake, named in recognition of its unusual shape that serves as a landmark for bush pilots. The unusual shape of the lake attracted the attention of geologists and, in 1965, evidence of shock metamorphism was discovered at the site (Fig. A-17).

Pilot Lake is roughly square with a diameter of about 5 km. Its outline is unusual for the area. The lake has a relatively flat bottom with depths of 50 to 70 m. With an estimated original diameter of about 6 km, the Pilot impact structure is assumed to be a heavily eroded, complex impact structure. On the basis of the isotopic analysis of one impact melt rock, the Pilot impact structure is generally dated at 445 ± 2 Ma.

Presqu'île

The Presqu'île impact structure (lat. 49°43′N; long. 74°84′W) is in central Quebec, about 10 km south of Chapais (Fig. A-1). It includes de la Presqu'île Lake, which is named for the large promontory on its eastern shore that extends into the lake, nealy forming an island. The impact structure was discovered during regional mapping.

De la Presqu'île Lake is a roughly ring-shaped lake about 6 km in diameter. The surrounding topography is one of rolling hills, and no indication has been found of a rim to the structure. The structure's diameter is poorly defined by the presence of shatter cones, which occur over an area 6 km in diameter. If the centre of the structure is assumed to be located on de la Presqu'île Lake, with the large eastern promontory representing the remnant of a heavily eroded central uplift, then the known distribution of shatter cones gives an

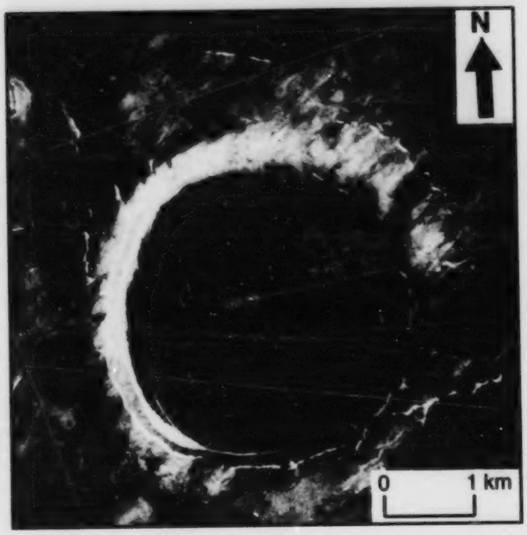


Figure A-16. *Nouveau Québec impact structure. National Airphoto Library photograph*

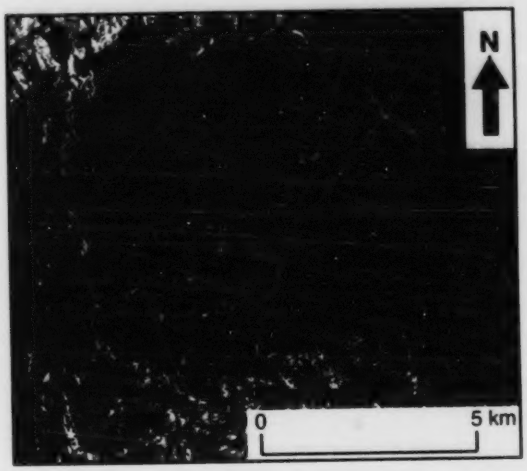


Figure A-17. *Pilot impact structure. National Airphoto Library photograph* estimated original diameter closer to 24 km. The age of the structure is unknown, but is generally considered to be less than 500 Ma.

Saint Martin

The Saint Martin impact structure (lat. 51°46'N; long. 98°32'W) is located in central Manitoba between lakes Manitoba and Winnipeg (Fig. A-1). It takes its name from Lake St. Martin, which lies towards its southeastern boundary. The area has been of considerable geological interest because of outcropping Precambrian granite and younger unmetamorphosed 'volcanic' and sedimentary rocks. An impact origin was first proposed from the analysis of rock samples drilled from the site.

The impact structure essentially has no surface expression. It is partly filled by up to 80 m of sedimentary rocks overlain by up to 30 m of glacial drift. A structurally uplifted area in the centre is about 22 km in diameter. This was considered originally to represent the diameter of the structure, but more recent interpretations suggest that it is the diameter of the central uplift area and surrounding ring-like trough. The limit of known structural disturbance, as defined by the drillhole data, is about 40 km, which is now the accepted estimate for the diameter of the structure, which thus is a complex structure with a central uplifted area. The impacting body may have been an achondritic (stony) body of variable composition. The age for the Saint Martin impact event is generally given as 219 ± 32 Ma, on the basis of the isotopic analysis of impact melt rocks.

Slate Islands

The Slate Islands impact structure (lat. 48°40'N; long. 87°00'W) is in Lake Superior, 13 km south of Terrace Bay, Ontario (Fig. A-1). Geological mapping of the islands in 1974 led to the discovery of extensive breccias and shatter cones.

The Slate Islands impact structure is almost totally submerged beneath the waters of Lake Superior. The only surface expression is a circular group of islands about 7 km in diameter (Fig. A-18) that rise relatively steeply out of the lake and are surrounded by a discontinuous submarine trough, about 15 to 16 km in diameter, where water depths reach over 250 m. The trough in turn is surrounded by a discontinuous ring-like ridge, where water depths rise to less than 100 m. The ridge is believed to approximate the original diameter of the structure at about 30 km. The overall form of the structure (rim, trough, and central peak) is that of a complex central-peak impact structure. No materials suitable for isotopic dating have been obtained from the structure. The levels of recorded shock and their variation in the rocks of the central uplift are comparable to those at Charlevoix, suggesting a similar level of erosion. On this basis, an age of less than 350 Ma is suggested for the Slate Islands structure.

Steen River

The Steen River impact structure (lat. 59°31'N; long. 117°39'W) is in northwestern Alberta, about 700 km northwest of Edmonton (Fig. A-1). It takes its name from the river that runs across its eastern boundary. The structure is completely buried by over 180 m of sedimentary rock and was first detected as an anomaly during magnetic and seismic surveys. Imperial Oil Enterprises drilled it in 1963 and encountered Precambrian basement rock some 760 m above the expected regional level and 'pyroclastic' (volcanic) rocks. Although the basement rocks were only partly cored, shock-metamorphic features were identified in 1966.

Since it cannot be seen at the surface, a general impression of the structure must be obtained from drillhole data and geophysical work. Its form is that of a complex central-peak structure, about 25 km in diameter, but the details of its morphology are not well known. In the centre, the basement forms a central peak whose base is approximately 8 km in diameter. The peak rises about 1700 m from the base of the ring-like trough in the southeast to about 760 m above regional basement rocks. A single isotopic age on a sample of impact melt rock dates the feature at 95 ± 7 Ma. Oil and gas

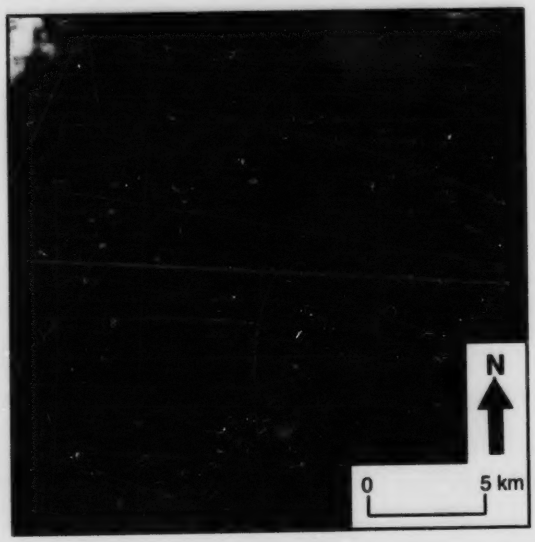


***Figure A-18.** Slate Islands impact structure. Landsat image courtesy of the National Aeronautics and Space Administration* shows are found within the Steen River structure, which is the site of exploration activity. Currently, there is minor oil production from the Steen River structure. However, there is every indication that it will become a significant natural gas play.

Sudbury

The Sudbury impact structure in central Ontario takes its name from the city of Sudbury, which lies within it (Fig. A-1). It is the largest and oldest impact structure in Canada and the site of world-famous nickel, copper, and platinum deposits, which are associated with the Sudbury Igneous Complex.

The most prominent feature of the Sudbury structure is the approximately 30 km by 60 km elliptical basin formed by the Sudbury Igneous Complex, the interior of which is known as the 'Sudbury Basin' (Fig. A-19). Neither the Sudbury Igneous Complex nor the Sudbury Basin are the same as the Sudbury impact structure, but are only part of the structure. The elliptical shape of the Sudbury Igneous Complex was caused by Precambrian (older than 570 Ma) tectonic deformation. Recent analyses suggest that the original diameter of the final rim of the structure was 250 km or more. Because of the tectonic deformation and the considerable erosion (estimated at about 10 km) that have taken place, the original form of the Sudbury impact structure is difficult to reconstruct. From its estimated original dimensions, it was most likely a peak-ring or possibly a multiring basin. On the basis of isotopic dating, the age of the Sudbury impact event is generally given as 1850 ± 3 Ma.

Viewfield

The Viewfield impact structure (lat. 49°35′N; long. 103°04′W) is located in southeastern Saskatchewan (Fig. A-1). It was first suggested as an impact structure in the early 1970s, on the basis of its form, as determined from geophysical work and drilling. At that time, however, no evidence of shock metamorphism was forthcoming.

The Viewfield structure has no surface expression and is completely buried, occurring at a depth of over 1.2 km. From geophysical work and hydrocarbon wells, the structure has the form of a simple impact structure with a diameter of about 2 to 3 km. Wells in the centre of the structure penetrate at least 400 m of breccia, with the degree of brecciation decreasing with depth. In the 1990s, a re-examination of cores from the wells at Viewfield confirmed the occurrence of microscopic planar deformation features in the mineral quartz. There is hydrocarbon production from the Viewfield impact structure. Oil production is from structures in the rim area and has been ongoing since 1969. There are approximately 50 producing oil wells, with primary reserves estimated at over 10 million barrels of oil and 4 billion cubic feet of natural gas. Most interestingly, the success ratio of hydrocarbon-producing to 'dry' holes at Viewfield is a remarkable 82%. On the basis of stratigraphic considerations, the Viewfield impact structure is estimated to be approximately 190 ± 20 Ma.

Wanapitei

The Wanapitei impact structure (lat. 46°45′N; long. 80°45′W) is in central Ontario, about 40 km northeast of the city of Sudbury (Fig. A-1). It lies entirely within Lake Wanapitei and within the original limits of the Sudbury impact structure, but is much younger (Fig. A-19). Attention was first drawn to the lake and its possible relation to an impact structure during a gravity survey of the Sudbury region in 1969. An examination of glacial till south of the lake led to the discovery of breccia boulders with abundant shock-metamorphic effects.

An area about 8 km in diameter in the northern part of the lake is devoid of islands. There is no topographic indication of the rim or of a central uplift area, presumably due to post-impact erosion. A central uplift, however, is generally assumed to have once existed. In general, the diameter of the structure is given as 7.5 km, which corresponds largely to the island-free, northern part of the lake. Geochemical studies of three impact melt rocks and target-rock quartzite indicate that

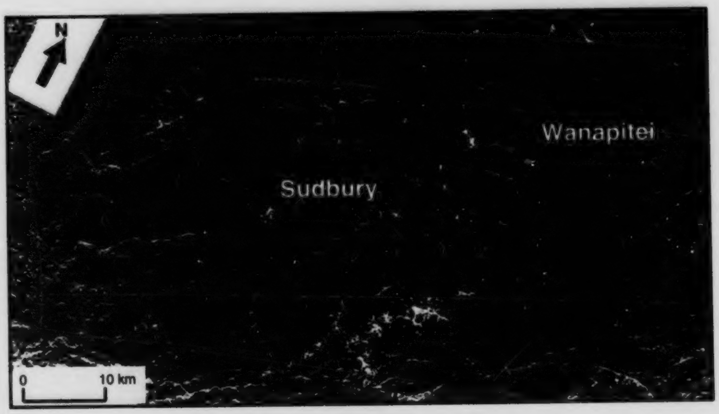


Figure A-19.

Sudbury and Wanapitei impact structures (radar image courtesy of the Earth Physics Branch).

the impacting body was a chondrite (rich in silicate minerals) meteorite. The Wanapitei structure is generally dated at 37 ± 2 Ma, on the basis of isotopic analyses of two impact melt rocks.

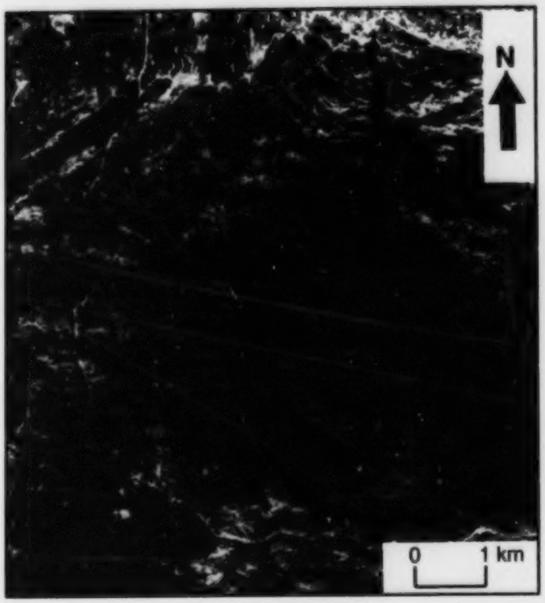


Figure A-20. West Hawk impact structure. National Airphoto Library photograph

West Hawk

The West Hawk structure (lat. 49°46′N; long. 95°11′W) is in southeastern Manitoba, close to the Ontario border (Fig. A-1). It is contained within the lake for which it is named (Fig. A-20). It was first noted as a possible impact site in 1960, following reconnaissance field studies and a gravity survey in 1958–1959. Additional gravity observations were conducted in 1962 and defined a circular negative gravity anomaly. Four drillholes were sunk off the ice in the winters of 1965 and 1966, and about 1 km of core was recovered. Breccia and brecciated target rocks were encountered, along with evidence of shock metamorphism.

West Hawk Lake is roughly circular with a diameter of 3.6 km. Compared to other lakes in the area, it is remarkably deep, reaching depths of 110 m. Drill data indicate that the West Hawk Lake structure has a simple, bowl-shaped form. By analogy with the Brent structure, the original diameter is estimated at 2.4 km. The age of the West Hawk impact event is tentatively estimated at 100 ± 50 Ma, on the basis of the stratigraphic relationship of the structure to underlying and overlying rocks.

Sea-level rise in Canada

J. Shaw[1], R.B. Taylor[2], D.L. Forbes[1], and S. Solomon[1]

Shaw, J., Taylor, R.B., Forbes, D.L., and Solomon, S., 2001: Sea-level rise in Canada; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 225–226.*

Abstract: Canadian coasts are threatened by a global sea-level rise of 49 cm by 2100, due to the effects of climate change. This contribution shows the sensitivity of the coasts of Canada to this increase. The largest region of high sensitivity comprises much of the coasts of Nova Scotia, Prince Edward Island, and New Brunswick. In the Arctic, the coast of the Beaufort Sea is highly sensitive. On the Pacific coast, areas of high sensitivity are the urbanized Fraser River delta and parts of Graham Island. The chief hazards associated with sea-level rise are the loss of property and infrastructure as a result of coastal erosion, and the damage to agricultural land, property, and infrastructure by flooding. Four examples are illustrated: Beaufort Sea; Placentia, Newfoundland; eastern shore, Nova Scotia; and head of the Bay of Fundy.

Résumé : Les côtes canadiennes sont menacées par une montée globale du niveau de la mer qui atteindra 49 cm en 2100 en raison des incidences du changement climatique. Le présent document met en évidence la sensibilité des côtes canadiennes à cette montée. La région la plus vaste et la plus sensible à ce phénomène englobe la presque totalité des côtes de la Nouvelle-Écosse, de l'Île-du-Prince-Édouard et du Nouveau-Brunswick. Dans l'Arctique, la côte de la mer de Beaufort est une région très sensible alors que sur la côte du Pacifique, les zones très sensibles sont les régions urbanisées du delta du Fraser et des portions de l'île Graham. Les principaux risques associés à la montée du niveau de la mer sont la perte de biens et d'infrastructures attribuable à l'érosion côtière et les dommages causés par les inondations aux terres agricoles, aux biens et aux infrastructures. Quatre exemples sont présentés, notamment la mer de Beaufort, Placentia (Terre-Neuve), la côte est de la Nouvelle-Écosse et le fond de la baie de Fundy.

---

[1] Geological Survey of Canada, 1 Challenger Drive, Dartmouth, Nova Scotia B2Y 4A2
[2] Mineral Resources Division, 601 Booth Street, Ottawa, Ontario K1A 0E8

BACKGROUND

Sea levels along Canada's coasts have been changing for thousands of years, principally as a result of glacio-isostasy and eustasy. In the future, Canada's coasts may undergo a relatively rapid change in sea level as a result of the effects of climate change. Figure 1 (in pocket) shows the boundary line separating submerging and emerging parts of the Canadian coast.

Environmental geochemistry and geochemical hazards

Compiled by R.D. Knight[1] and R.A. Klassen[1]
with contributions by D.R. Boyle[2], J.M. Carson[2], B.W. Charbonneau[2], S.J.A. Day[2], P.W.B. Friske[2], R.G. Garrett[2], G.E.M. Hall[2], P. Henderson[1], R.J. Hetu[2], I.M. Kettles[1], R.A. Klassen[1], M.W. McCurdy[2], I. McMartin[1], A. Plouffe[1], P.E. Rasmussen[2], W.W. Shilts[1], and L.H. Thorleifson[1]

Knight, R.D. and Klassen, R.A. (comp.), 2001: Environmental geochemistry and geochemical hazards; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 227–230.*

Abstract: Environmental geochemistry is concerned with the abundance, distribution, and mobility of elements at the Earth's surface. It is important for land use management and health protection because Earth surface geochemistry affects life. Through regional surveys and detailed studies, the Geological Survey of Canada (GSC) provides the key geoscience basis for managing environmental issues. This work shows the natural distributions and concentrations of elements for much of Canada. It tells us about their bedrock origins, mineral hosts, and chemical forms, and the dispersal processes associated with their distribution in the landscape. Natural background variations shown by GSC maps can be used to identify areas of potential geochemical hazard and to distinguish between natural and industrial sources. This synthesis shows the application of geoscience research to environmental geochemistry.

Résumé : La géochimie environnementale traite de l'abondance des éléments à la surface de la Terre, de leur répartition et de leur mobilité. Elle joue un rôle important dans l'aménagement du territoire et la protection de la santé parce que la géochimie de la surface du globe a une incidence sur les organismes vivants. Les levés effectués par la Commission géologique du Canada (CGC) à l'échelle régionale ainsi que les études détaillées fournissent d'importantes données géoscientifiques de base qui servent à la gestion des problèmes environnementaux. L'article présente la répartition et la concentration naturelles des éléments dans une grande partie du Canada. Il nous renseigne sur leur origine dans le substratum rocheux, leurs minéraux hôtes et leurs formes chimiques, ainsi que les processus de dispersion associés à leur répartition dans le paysage. Les variations de fond naturelles mises en évidence sur les cartes de la CGC permettent d'identifier les régions présentant des risques géochimiques potentiels et de différencier les sources naturelles des sources industrielles. Cette synthèse montre les applications de la recherche géoscientifique à la géochimie environnementale.

---

[1] Terrain Sciences Division, Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

[2] Mineral Resources Division, Mineral and Regional Geoscience Branch, 601 Booth Street, Ottawa, Ontario K1A 0E8

BACKGROUND

The purpose of this synthesis is to summarize thematic aspects of environmental geochemistry that relate to hazards. Emphasis is placed on a small group of elements that, in abnormal concentrations (high or low), can represent a health risk. Contributions to this synthesis were selected to show a range of potential geochemical hazards throughout Canada that can occur in different types of surficial materials, groundwater, and lakes, and the geological processes affecting them. These contributions are presented in Figure 1 (in pocket).

SUGGESTED READING

Groundwater geochemistry and environmental issues

Boyle, D.R.

1991: The Canadian geochemical environment and its relationship to the development of health status indicators; *in* Environmental Health Status Indicators, (ed.) R.G. McColl; University of Waterloo Press, Waterloo, Ontario, p. 75-115.

Boyle, D.R. and Chagnon, M.

1995: Fluoride toxicity from groundwaters in the Maria Area of the Maritime Carboniferous Basin, Gaspe Region, Quebec; Journal of Environmental Geochemistry and Health, v. 17, p. 5–12.

Boyle, D.R., Spirito, W.A., and Adcock, S.W.

1994: Groundwater hydrogeochemical survey of Southeastern New Brunswick; Geological Survey of Canada, Open File 2912, 126 p.

1996: Groundwater hydrogeochemical survey of central New Brunswick; Geological Survey of Canada, Open File 3306, 89 p.

A geological perspective on the effects of acid rain on the environment

Kettles, I.M. and Shilts, W.W.

1991: Glacial geology and glacial sediment geochemistry in the Clyde Forks-Westport area of Ontario; Geological Survey of Canada, Paper 91-17, 34 p.

1994: Composition of glacial sediments in Canadian Shield terrane, southeastern Ontario and southwestern Quebec: applications to acid rain research and mineral exploration; Geological Survey of Canada, Bulletin 463, 33 p

1995: Geochemical and lithologic composition of surficial sediments, southeastern Ontario; Geological Survey of Canada, Open File 3175, 58 p.

Transport of mercury by glaciers — central British Columbia

Plouffe, A.

1995: Glacial dispersal of mercury from bedrock mineralization along Pinchi Fault, north central British Columbia; Water, Air, and Soil Pollution, v. 80, p. 1109–1112.

1995: Geochemistry, lithology, mineralogy and visible gold grain content of till in the Manson River and Fort Fraser map areas, central British Columbia (NTS 93K and N); Geological Survey of Canada, Open File 3194, 119 p.

1997: Physical partitioning of mercury in till: an example from British Columbia; Journal of Geochemical Exploration, v. 59, p. 219–232.

1998: Detrital transport of metals by glaciers, an example from the Pinchi Mine site, central British Columbia; Environmental Geology, p. 183–196.

The effects of smelter emmisions on the terrestrial environment

Bonham-Carter, G.F. and McMartin, I.

1997: Calculations on metals in humus around the base metal smelter at Flin Flon, Manitoba, Canada; *in* Proceedings, Third Annual Conference of International Association for Mathematical Geology (IAMG'97), (ed.) V. Pawlowsky-Glahn, p. 262–269.

Henderson, P.J.

1995: Surficial geology and drift composition of the Annabel Lake-Amisk Lake area, Saskatchewan (NTS 63L/9, L/16 and part of 63K/12 and K/13); Geological Survey of Canada, Open File 3026.

Henderson, P. and McMartin, I.

1995: Mercury distribution in humus and surficial sediments in the vicinity of Flin Flon, Manitoba, Canada; Water, Air, and Soil Pollution, v. 80, p. 1043–1046.

Henderson, P.J., McMartin, I., Hall, G.E.M., Walker, D., and Percival, J.B.

1998: The geochemical and physical characteristics of heavy metals in humus and till in the vicinity of the base metal smelter at Flin Flon, Manitoba; Environmental Geology, p. 39–58.

McMartin, I. and Henderson, P.J.

1996: Regional distribution of heavy metals in humus and till near a base metal smelter at Flin Flon, Manitoba, Canada; EXPLORE (Newsletter for the Association of Exploration Geochemists), no. 93, October, p. 10–11.

McMartin, I., Henderson, P.J. and Nielsen, E.

1999: Impact of a base metal smelter on the geochemistry of soils of the Flin Flon region, Manitoba and Saskatchewan; Canadian Journal of Earth Sciences, p. 141–160.

McMartin, I., Henderson, P.J., Nielsen, E., and Campbell, J.E.

1996: Surficial geology, till and humus composition across the Shield Margin, north-central Manitoba and Saskatchewan; geospatial analysis of glaciated environment; Geological Survey of Canada, Open File 3277, 300 p.

Naturally occurring cadmium in upland soils and lake-bottom sediments of the boreal forest

Hall, G.E.M., Gauthier, G., Pelchat, J.C., Pelchat, P., and Vaive, J.E.

1996: Application of a sequential extraction scheme to ten geological standard reference materials for the determination of 20 elements; Journal of Analytical Atomic Spectrometry, v. 11, p. 787–796.

Hall, G.E.M., Vaive, J.E., Beer, R., and Hoashi, M.

1996: Selective leaches *revisited*, with emphasis on the amorphous Fe oxyhydroxide phase extraction; Journal of Geochemical Exploration, v. 56, p. 59–78

Rasmussen, P.E.,

1991: Lake sediment sampling in the Huntsville area — The distribution of Hg and other elements in deep sediments of the Huntsville lakes: a preliminary study; Ontario Geological Survey, Open File Report 5818, 86 p.

1996: Trace metals in the environment: A geological perspective; Geological Survey of Canada, Bulletin 429, 26 p.

Prairie soil geochemistry

Garrett, R. G.,

1994: The distribution of cadmium in A-horizon soils in the praries of Canada and adjoining United States; *in* Current Research 1994-B; Geological Survey of Canada, p. 73–82

Environmental applications of gamma ray spectrometry surveys

Canada Mortgage and Housing Corporation

1997: Radon: A Guide for Canadian Homeowners; Canada Housing and Mortgage Corporation, 40 p.

Charbonneau, B.W

1991: Geophysical signature, geochemical evolution and radioactive mineralogy of the Fort Smith radioactive belt, Northwest Territories, Canada; in Primary Radioactive Minerals (The Textural Patterns of Radioactive Mineral Paragenetic Associations); Theophrastuss Publications, Athens, Greece, p. 21–48.

Charbonneau, B.W., Killeen, P.G., Carson, J.M., Cameron, G.W., and Richardson, K.A.

1976: The significance of radioelement concentration measurements made by airborne gamma-ray spectrometry over the Canadian Shield; IAEA Symposium on Exploration for Uranium Ore Deposits, Vienna, Austria, 1976, STI/PUB/434, p. 35–53.

Charbonneau, B.W., Richardson, K.A., and Grasty, R.L.

1973: Airborne gamma-ray spectrometry as an aid to geological mapping Township No. 155 Elliot Lake area, Ontario; *in* Report of Activities, Part B; Geological Survey of Canada, Paper 73-1B, p. 39–47.

Darnley, A.G. and Grasty, R.L.

1971: Airborne radiometric survey of the Elliot Lake area, Ontario (part of NTS 41J); Geological Survey of Canada, Open File 75, scale 1:50 000.

1972: Airborne radiometric survey of the Fort Smith area, NWT; NTS 74M, 75D, 75E; Geological Survey of Canada, Open File 101, scale 1:250 000.

Darnley, A.G., Richardson, K.A., Grasty, R.L., Carson, J.M., Holman, P.B., and Charbonneau, B.W.

1986: Radioactivity map of Canada (first edition); Geological Survey of Canada, Map 1600A, scale 1:5 000 000.

Eisenbud, M.

1977: Summary report; *in* International Symposium on Areas of High Natural Radioactivity, (ed.) Academia Brasileira de Ciências, Rio de Janeiro, p. 167–179.

Ford, K.L., Carson, J.M., Grant, J.A., and Holman, P.B.

1989: Radioactivity map of Nova Scotia: uranium (ppm); Geological Survey of Canada, Map 35006G, scale 1:500 000.

Gandhi, S.S.

1995: An overview of the exploration history and genesis of Proterozoic uranium deposits in the Canadian Shield; Exploration and Research for Atomic Minerals, v. 8, p. 1–47.

Grasty, R.L., Carson, J.M., Charbonneau, B.W., and Holman, P.B.

1984: Natural background radiation in Canada; Geological Survey of Canada, Bulletin 360, 39 p.

Jackson, S.A.

1992: Estimating radon potential from an aerial radiometric survey; Health Physics, The Radiation Protection Journal, v. 62, no. 5, p. 450–452.

Killeen, P.G.

1979: Gamma ray spectrometric methods in uranium exploration — application and interpretation; *in* Geophysics and Geochemistry in the Search for Metallic Ores, (ed.) P.J. Hood; Geological Survey of Canada, Economic Geology Report 31, p. 163–229.

Richardson, K.A., Holman, P.B., and Elliott, B.E.

1975: Airborne radiometric survey of the Blind River Area, Ontario, NTS 41J; Geological Survey of Canada, Open File 262, scale 1:250 000.

Richardson, K.A., Holman, P.B., and Hyatt, W.G.

1977: Airborne gamma-ray spectrometry survey, Elliott Lake area (part of 41J); Geological Survey of Canada, Open File 502, scale 1:20 000.

Shives, R.B.K., Ford, K.L., and Charbonneau, B.W.

1995: Applications of gamma ray spectrometry magnetic/VLF-EM surveys 'Workshop Manual'; Geological Survey of Canada, Open File 3061, 82 p.

Mercury in lake and stream sediments

Cameron, E.M. and Ballantyne, S.B.

1977: Reconnaissance-level geochemical and radiometric exploration data from the vicinity of the Rabbit Lake uranium deposit; The Canadian Institute of Mining and Metallurgy Bulletin, v. 70, p. 76–85.

Coker, W.B. and Shilts, W.W.

1979: Lacustrine geochemistry around the north shore of Lake Superior: implications for evaluation of effects of acid precipitation; Geological Survey of Canada, Paper 79-1C, p. 1–15.

Coker, W.B., Hornbrook, E.H.W. and Cameron, E.M.

1979: Lake sediment geochemistry applied to mineral exploration; *in* Geophysics and Geochemistry in the Search for Metallic Ores; (ed.) P.J. Hood; Geological Survey of Canada, Economic Geology Report 31, p. 435–477.

Davenport, P.H., Christopher, T.K., Vardy, S., and Nolan, L.W.

1992: Geochemical mapping in Newfoundland and Labrador: its role in establishing geochemical baselines for the measurement of environmental change; *in* Current Research, Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 92-1, p. 281–296.

Fortescue, J.A.C. and Vida, E.Al

1990: Geochemical survey, Hanes Lake area; Ontario Geological Survey, Map 80 806, scale 1:50 000.

Friske, P.W.B.

1985: Regional geochemical reconnaissance: interpretation of data from the north shore of Lake Superior; Geological Survey of Canada, Paper 84-21, 38 p.

1991: The application of lake sediment geochemistry in mineral exploration; Geological Survey of Canada, Open File 2390, Paper no. 4, p. 4-1 to 4-20.

Friske, P.W.B. and Coker, W.B.

1995: The importance of geological controls on the natural distribution of mercury in lake and stream sediments across Canada; Water, Air, and Soil Pollution, v. 80, p. 1047–1051.

Friske, P.W.B. and Hornbrook, E.H.W.

1991: Canada's National Geochemical Reconnaissance programme; Transactions of the Institution of Mining and Metallurgy, Section B, v. 100, p. 47–56.

Friske, P.W.B., McCurdy, M.W., Day, S.J.A., and Adcock, S.W.

1994: National Geochemical Reconnaissance — Ontario compilation; Distribution of copper in 17 934 lake sediment samples; Geological Survey of Canada, Open File 2959, scale 1:50 000.

Garrett, R.G.

1969: The determination of sampling and analytical errors in exploration geochemistry; Economic Geology, v. 64, p. 568–569.

1974: Field data acquisition methods for applied geochemical surveys at the Geological Survey of Canada; Geological Survey of Canada, Paper 74-52, 36 p.

Garrett, R.G., Banville, R.M.P. and Adcock, S.W.

1990: Regional geochemical data compilation and map preparation, Labrador, Canada; *in* Regional Geochemical Mapping, (ed.) A.G. Darnley and R.G. Garrett; Journal of Geochemical Exploration, v. 39, p. 91–116.

Hornbrook, E.H.W. and Garrett, R.G.

1976: Regional geochemical lake sediment survey east-central Saskatchewan; Geological Survey of Canada, Paper 75-41, 20 p.

Johnson, M.G., Culp, L.R., and George, S.E.

1986: Temporal and spatial trends in metal loadings to sediments of the Turkey Lakes, Ontario; Canadian Journal of Fisheries and Aquatic Sciences, v. 43, p. 754–762.

Jonasson, I.R.

1976: Detailed hydrogeochemistry of two small lakes in the Grenville geological province; Geological Survey of Canada, Paper 76-13, 37 p.

Kerr, A. and Davenport, P.H.

1990: Application of geochemical mapping techniques to a complex Precambrian shield area in Labrador, Canada; *in* Regional Geochemical Mapping, (ed.) A.G. Darnley and R.G. Garrett; Journal of Geochemical Exploration, v. 39, p. 225–247.

Lynch, J.J.

1990: Provisional elemental values for eight new geochemical lake sediment and stream sediment reference materials LKSD-1, LKSD-2, LKSD-3, LKSD-4, STSD-1, STSD-2, STSD-3 and STSD-4; Geostandards Newsletter, v. 14, no. 1, p. 153.

Maurice, Y.T.

1977: Geochemical methods applied to uranium exploration in southwest Baffin Island; The Canadian Institute of Mining and Metallurgy Bulletin, v. 70, p. 96–103.

Ramsey, M.H., Thompson, M., and Hale, M.

1992: Objective evaluation of precision requirements for geochemical analysis using robust analysis of variance; Journal of Geochemical Exploration, v. 44, p. 23–36.

Rogers, P.J., Chatterjee, A.K. and Aucott J.W.
1990: Metallogenic domains and their reflection in regional lake sediment surveys from the Meguma Zone, southern Nova Scotia, Canada; *in* Regional Geochemical Mapping, (ed.) A.G. Darnley and R.G. Garrett; Journal of Geochemical Exploration, v. 39, p. 153–174.

Glaciation, bedrock composition, and trace metals in the environment

Klassen, R.A.
1993: A preliminary interpretation of ice flow patterns and glacial history, central Newfoundland (extended abstract); *in* Report of Activities, 1993; Geological Survey Branch, Government of Newfoundland and Labrador, Department of Mines and Energy, p. 34–40.
1994: A preliminary interpretation of ice flow patterns and glacial history, central Newfoundland; *in* Current Research, 1994-D; Geological Survey of Canada, p. 13–22.
1998: Geological factors affecting the distribution of trace metals in glacial sediments of central Newfoundland; Environmental Geology, v. 33, no. 2/3, p. 154–169.

Klassen, R.A. and Murton, J.B.
1996: Quaternary geology of the Buchans area, Newfoundland: implications for mineral exploration; Canadian Journal of Earth Sciences, v. 33, no. 2, p. 363–377.

Eolian activity

S.A. Wolfe[1]

Wolfe, S.A., 2001: Eolian activity; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 231–240.*

Asbtract: Eolian processes pertain to the transport of sediment by wind. This paper reviews eolian activity in Canada, and eolian activity as a geological hazard. The effects of wind erosion can produce long-term impacts or damage to agricultural and recreational lands, and, in some cases, to human health. The hazards include wind erosion of agricultural soils, dust storms, sand-dune activity caused by human disturbance, overgrazing, and/or drought, deflation areas, and alkali dust emissions. Many such hazards result from human activities that initiate or accelerate the rate of eolian processes. Although wind erosion is currently under control in most regions of Canada, projections of future climate warming and increased aridity in some regions suggest that the risk of wind erosion and sand-dune activity could increase in future years.

Résumé : Les processus éoliens sont liés au transport des sédiments par le vent. Le présent article passe en revue l'activité éolienne au Canada et les risques géologiques qu'elle engendre. Les incidences de l'érosion éolienne sont susceptibles d'avoir des répercussions à long terme ou d'endommager les terres agricoles et les terres de loisirs et, dans certains cas, de nuire à la santé des humains. Les risques sont, entre autres, l'érosion des sols agricoles, les tempêtes de poussière et l'édification et la destruction de dunes engendrées par l'activité humaine, le surpâturage et la sécheresse, la formation de zones de déflation et les émissions de poussière alcaline. Bon nombre de ces risques sont liés à l'activité humaine qui déclenche les processus éoliens ou en accélère le rythme. Bien que l'érosion éolienne soit actuellement enrayée dans la plupart des régions du Canada, le réchauffement du climat dans l'avenir et l'augmentation de l'aridité dans certaines régions permettent de supposer que les risques liés à l'érosion éolienne et à l'édification et la destruction des dunes dans les années à venir sont susceptibles de s'accroître.

---

[1] Terrain Sciences Division, Geological Survey of Canada, Ottawa, Ontario

INTRODUCTION

What are eolian processes?

Eolian processes involve the movement of sediment by wind. Sand is transported by wind typically no more than few metres above the ground, in a bouncing-and-hopping motion called 'saltation'. Silt and clay grains are commonly carried aloft, or 'suspended', in the air, and may rise many hundreds of metres or more into the atmosphere (Fig. 1). By transporting and depositing sediment, eolian activity may result in wind erosion and the formation of sand dunes. It can also cause dust storms, during which fine soil particles are entrained into the air and redeposited great distances away.

Eolian processes require both wind and a supply of erodible sediment. Most areas of Canada are sufficiently windy so that fine-grained sediment can be eroded and transported, and are in fact windier than most of the 'sand-sea' deserts of the world, including those of Australia, Saudi Arabia, and California. However, because Canada has a comparatively humid climate, the fine-grained sediment that could potentially be blown by wind is covered by vegetation or bound by moisture. For this reason, eolian activity in Canada is not as pervasive as in more arid regions of the world. Nevertheless, wind erosion, dust storms, and sand dunes occur throughout Canada, at times resulting in hazards to Canadians.

EOLIAN ACTIVITY IN CANADA

The accompanying map of eolian activity in Canada (Fig. 2, in pocket) depicts the locations and relative activity of sand dunes, dust storms, and risk of wind erosion in Canada. Sand-dune areas greater than 25 km^2 are outlined in detail, and smaller dune areas are represented by individual dots. Sand dunes are found in both coastal and inland settings throughout Canada. Most inland dunes are derived from sand originally deposited in glacial lakes or rivers during waning stages of the last glaciation (14 000–8000 BP) and subsequently reworked by wind into dunes. Stabilized sand dunes are covered by vegetation and are not currently affected by wind erosion. Partly active dune areas are primarily stabilized by vegetation, but include some zones of small, active sand dunes or blowouts, which are depressions or hollows formed by wind erosion. Widespread active dune areas are dominated by active sand dunes forming a continuous surface of wind-blown sand. Coastal sand dunes represent areas of marine and lakeshore dunes. Deflation plains are areas of active wind erosion, but without sand dunes.

Figure 2 (inset) also indicates the risk of wind erosion to bare, unprotected agricultural soils in Canada. The risk of wind erosion to bare, unprotected soils is largely a function of soil texture, wind regime, and surface soil moisture conditions. In addition to these inherent conditions, agricultural practices and vegetation cover modify the risks of erosion. The vegetation cover acts to reduce wind erosion, such that on well vegetated surfaces, wind erosion is not a problem. Agriculture, however, can intensify or reduce the risk of wind erosion depending upon particular tillage and cropping practices. The risk of wind erosion in Canada is greatest in the southern regions of the Prairie Provinces and in parts of southern Ontario (Fig. 2).

Dust storms are perhaps one of the most striking examples of wind erosion. They are caused by turbulent winds that entrain fine-sized particles into the air, thereby reducing visibility to less than 1000 m. Dust storms may form clouds or 'walls' of dust that transport tonnes of soil in a single event. Figure 2 shows the frequency of dust storms on the southern Prairies, where most dust storms in Canada occur.

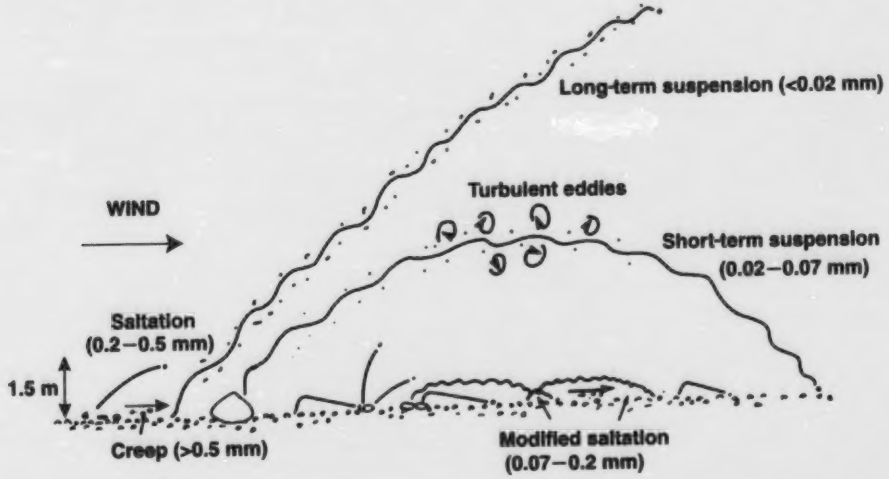

Long-term suspension (<0.02 mm)
Turbulent eddies
WIND
Short-term suspension (0.02–0.07 mm)
Saltation (0.2–0.5 mm)
1.5 m
Creep (>0.5 mm)
Modified saltation (0.07–0.2 mm)

Figure 1. Sediment transport by wind as a function of grain size (from Wolfe and Nickling, 1997).

Figure 2 is intended to raise public awareness of the widespread occurrence of eolian features and processes in Canada. In some areas, sand-dune deposits represent potential resources as aggregate material, grazing land, and forested regions. Elsewhere, agricultural, urban, recreational, and industrial development occurs on soils that are highly susceptible to wind erosion. Such developmen should be sustainable and be carried out in a way that minimizes the risk of wind erosion.

Eolian activity as a hazard

Collectively, eolian processes represent a hazardous condition whereby the effects of wind erosion can produce long-term impacts or damage. Many of these hazards result from human activities that initiate or accelerate the rate of eolian processes. An example is wind erosion of agricultural soils resulting in crop damage (by sandblasting) and in a loss of organic matter and soil nutrients. In extreme cases, dust storms can sweep across agricultural soils, causing reduced visibility and loss of nutrient-rich soil material. In other areas, hazards include erosion caused by the disturbance of stabilized and partly active sand dunes. In the Atlantic Provinces and Great Lakes regions, the recreational disturbance of coastal dunes has contributed to shoreline destabilization and erosion. In the dry, subhumid regions of the Prairies, stabilized and partly active sand-dune areas may be reactivated if the vegetation cover is reduced by drought, grazing, fire, or off-road-vehicle traffic. In other areas, including southern Ontario and boreal forest regions, deforestation and stripping of vegetation can result in deflation and reactivation of stabilized sand dunes if these dunes are not carefully managed or revegetated. In general, eolian processes have received relatively little attention in Canada. This disregard can lead to negative environmental and economic impacts resulting from eolian activity.

DISTRIBUTION OF EOLIAN ACTIVITY IN CANADA

Wind erosion of agricultural soils

Although agriculture is a necessity, providing both employment and a large proportion of the gross national product, it can also be viewed as representing one of the most extensive anthropogenic disturbances of soils in Canada. In most regions of Canada, soil erosion by water exceeds that by wind. However, wind erosion of agricultural soils is most severe and extensive when bare fields under low soil- moisture conditions are left exposed to wind. The greatest risk of wind erosion of agricultural soils is in the southern Prairie Provinces and in parts of southern Ontario (inset, Fig. 2). In the southern Prairie Provinces, the annual soil loss due to wind erosion has been estimated to be one third greater than that due to water (Fig. 3). Farming practices play a major role in determining the severity of wind erosion. For example, techniques that retain crop residue and reduce tillage have significantly reduced soil erosion by both wind and water. Soil-erosion risk has been mapped in Ontario and the Prairie Provinces by Agriculture Canada, and these maps are available to the public. By increasing public awareness of soils at risk of erosion, and by continuing to develop new soil conservation techniques, soil scientists and farmers are reducing the total agricultural land area affected by wind erosion in Canada.

Dust storms

Dust storms are one element of eolian processes that can be catastrophic, causing nearly instantaneous loss of crops and farm productivity. For the most part, however, the impacts are cumulative and can become costly over time. Dust storms cause soil and crop damage through erosion, burial, and sandblasting, the cumulative effects of which are decreased yields

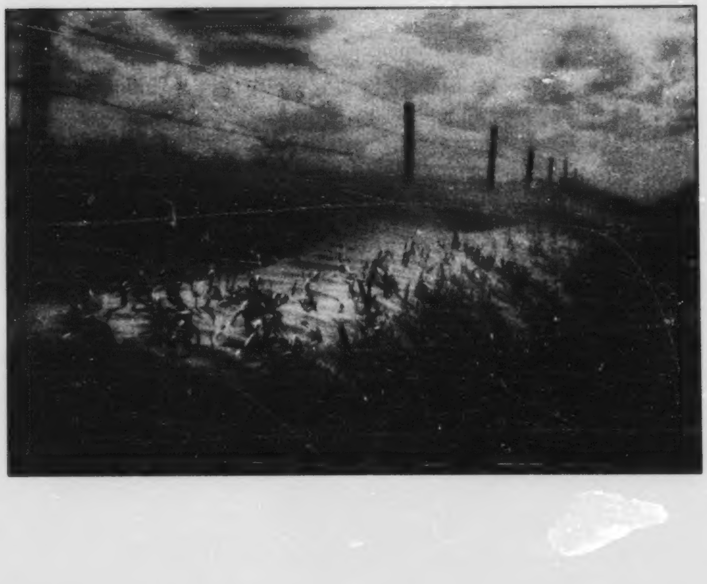


Figure 3.

Blown soil accumulating along a fence line south of Saskatoon, Saskatchewan, May 1989. Although modern farming practices do much to reduce the loss of soil by wind erosion, valuable soil is still eroded every year. Soil loss due to wind erosion in the Prairie Provinces is conservatively estimated to be about 160 tonnes per year (Sparrow, 1984). In this photograph, fine, nutrient-rich soil has been eroded by wind from a farm field and deposited along a fence line and ditch leaving only the coarser, less fertile soil on the field. Photograph by E. Wheaton, GSC 1995-069 and increased use of fertilizers. Dust storms also affect air and water quality, resulting in impacts on human health. In the United States and other countries, dust storms are considered significant hazards; however, in Canada, they have remained largely neglected. Consequently, few studies and little monitoring have been undertaken, and the overall impact of dust storms on human health, agriculture, transportation systems, and the economy is difficult to estimate.

In Canada, the occurrence of dust storms is closely related to agricultural activity on dry, erodible soils. During the 1930s, a series of dry years across the Great Plains of the United States and Canada resulted in what later became known as the 'dust bowl' and the 'Dirty Thirties'. Soil drifting and blowing was a serious problem from 1933 to 1938. In 1937, the total annual precipitation at Swift Current, Saskatchewan, amounted to only 21 cm, temporarily placing the area in the realm of a 'desert' (less than 25 cm of annual precipitation). Dust storms blew nutrient-rich soils off agricultural fields and filled roadside ditches and fence lines. The events of the 1930s were greatly exacerbated by poor farming methods.

Today, windbreaks, trash-cover farming, strip-row cropping, and other techniques are used to reduce blowing soil on farmland. Despite the diligent soil management practices of most farmers, dust storms still occur on the Canadian Prairies (Fig. 4). Figure 2 shows the total number of dust-storm occurrences on the southern Prairies, where dust storms in Canada are most common, between 1977 and 1989. Dust storms are most common in spring, particularly in April, when winds are strong, precipitation is low, and vegetative cover is sparse. The greatest number of dust storms occur near Regina and Saskatoon, where typically more than five dust storms are

Figure 4.

Dust storm in southern Alberta, spring 1988. Although many Canadians associate dust storms with the 'dust bowl' of the 1930s, dust storms occur frequently on the Canadian Prairies, most commonly in the spring and late summer. They generally last only a few hours, but may be accompanied by a 'wall of dust'. In more technical terms, dust storms are defined as wind events accompanied by blowing soil that reduces visibility at eye level to 1 km or less. The frequency and severity of dust storms increase during times of drought, with nearly 40 dust storms reported in southern Saskatchewan during the 1988 drought and over 60 in 1981. Photograph courtesy of the Prairie Farm Rehabilitation Association

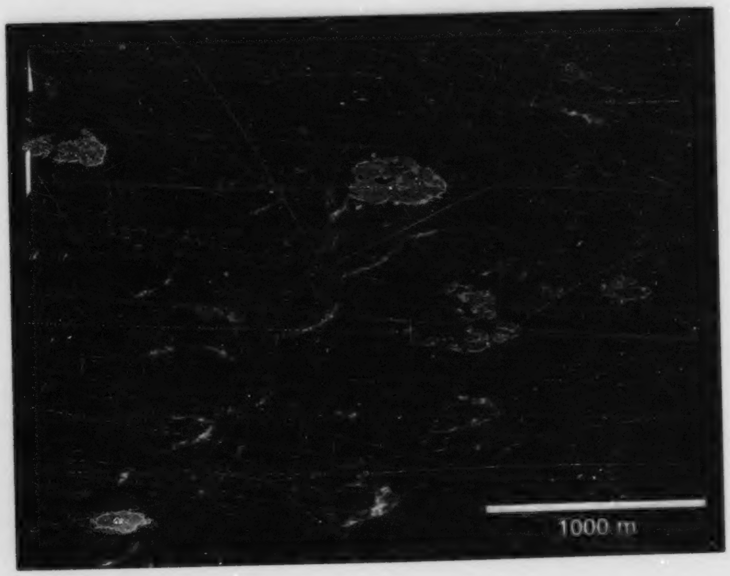


Figure 5.

Airphoto of stabilized sand dunes southeast of Grande Prairie, Alberta. Most sand dunes in Canada are currently stabilized by vegetation. The arms of these parabolic dunes point upwind, indicating that the prevailing wind was from west to east when the dunes were active. The sand dunes in the Grande Prairie region are thought to have been active about 13 000 years ago. NAPL A8919-17

Figure 6.

Sand dune encroaching on a schoolyard in Carcross, Yukon territory, June 1996. Due to intense disturbance along a pathway leading to the shore of Bennett Lake, this sand dune has migrated away from the beach towards the school. The migration of the dune is enhanced by strong winds blowing down the valley and across the lake. Other active dunes occur throughout the 'Carcross Desert' and many houses are built on stabilized sand dunes. GSC 1998-009A

recorded annually. In most areas, however, dust-storm frequency is underestimated because of a lack of reporting stations. Unless both monitoring and preventative steps are taken, dust storms will remain a costly, but neglected, hazard in Canada.

Sand dunes and deflation surfaces

Stabilized dunes

Stabilized sand dunes are the most common form of dunes in Canada (Fig. 2). Many were probably active immediately after deglaciation, in relatively arid environments with strong, off-ice (katabatic) winds. Eventually, most were stabilized by vegetation, as climate moderated and winds subsided. Currently, stabilized sand dunes occur in a wide range of environments including prairie grasslands, boreal forests, and arctic tundra (Fig. 5).

Disturbance of stabilized dunes is common, as many are utilized as sources of sand. In most instances, disturbance of these areas does not result in significant eolian activity or erosion, either because of adequate vegetation cover and soil-moisture content, or insufficient wind conditions. In some areas, however, localized erosion has been caused by deforestation, off-road-vehicle traffic, or simply intense trampling (Fig. 6). In most instances, major disturbances are due to forestry, agriculture, and sand mining on stabilized dunes. In the 1930s and 1940s, the deforestation of stabilized sand dunes in the Oak Ridges Moraine area of southern Ontario resulted in localized wind erosion and the need for reforestation.

Stabilized sand dunes have become active, most commonly as a result of inappropriate land use where the sensitivity of the dunes was not recognized. Consequently, the identification of stabilized sand dunes and their sensitivity to disturbance is the most important step to understanding potential hazards in these areas.

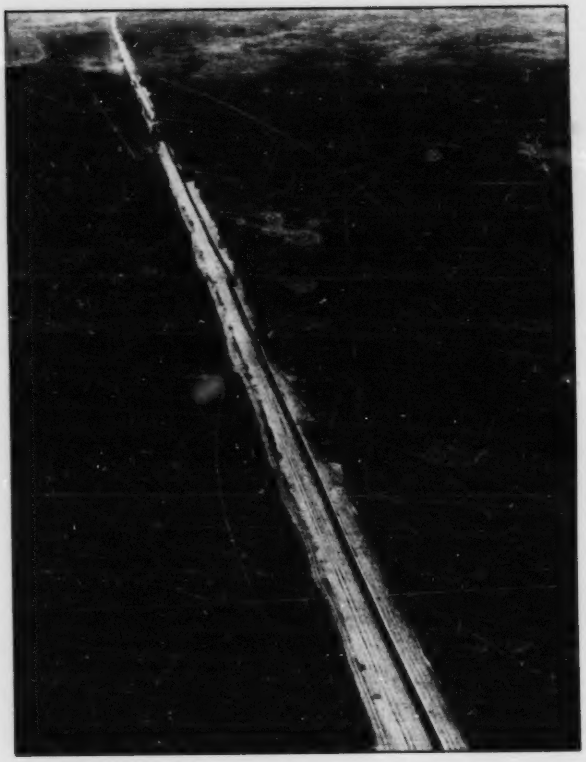


Figure 7. *Installation of a new line through stabilized sand dunes along the Trans-Canada Pipeline right-of-way southeast of Sceptre, Saskatchewan, September 1994. The Trans-Canada Pipeline crosses some of the most arid land on the Canadian Prairies, as well as stabilized sand dunes. After construction, the disturbed area must be reseeded to minimize the risk of wind erosion. GSC 1998-009C*

Partly active dunes

Partly active sand dunes are most susceptible to increased activity when the protective vegetative cover is reduced. Increased activity of inland dune areas may result from human disturbance and/or drought. Currently, partly active inland dunes are found primarily in southern Alberta, Saskatchewan, and Manitoba (Fig. 2).

The Great Sand Hills of southern Saskatchewan, which cover over 1000 km², are the largest partly active dune area in Canada. In this area, off-road-vehicle activity had disturbed vegetation in sand-dune areas. With the formation of a local planning commission, such activity was prohibited in sensitive areas and is closely monitored elsewhere (Fig. 7). Similarly, cattle ranching in the Great Sand Hills is conducted with strict regards to the carrying capacity of the land, so as to avoid overgrazing.

Partly active sand dunes on the southern Prairies are also susceptible to reactivation caused by drought (Fig. 8). Many sand-dune areas of southern Saskatchewan have been active in the last 200 years, and a future period of sustained drought (about 10 years) could result in widespread dune activity in the region. Furthermore, the southern prairies of Alberta and Saskatchewan, known as the 'Palliser Triangle' region, are more at risk of increased wind erosion as a result of climatic warming than any other region of Canada. Under such conditions, sand dunes would become even more sensitive to reactivation by disturbance. Consequently, existing measures to prevent dune activity may not be adequate for periods of prolonged drought.

Widespread active dunes

Only a few areas of widespread sand dune activity are found in Canada (Fig. 2). The dunes on the southern shores of Lake Athabasca are the largest occurrence in Canada (Fig. 9), covering over 4500 km². Within this area, active dunes cover

Figure 8.

Sand dunes west of Swift Current, Saskatchewan, April 1994. Active sand dunes such as this one are found in the drier prairie regions of southwestern Saskatchewan and southeastern Alberta. Many more sand dunes are stabilized by a thin cover of grass and soil. The sparse vegetation cover on these dunes makes them sensitive to changes in climate and to disturbance by animals and humans. During the droughts of the 1930s and the 1980s many of these sand dunes showed increased activity as the soils dried out. GSC 1998-009B

Figure 9.

Active sand dunes in the central part of the east dune field of the Athabasca Sand Dunes. The sand dunes of Lake Athabasca in northern Saskatchewan and Alberta represent the largest region of sand dunes in Canada. Within this enormous field of stabilized dunes is an area of active sand covering about 50 km². These active dunes present a striking contrast to the boreal forest cover that surrounds them. The dunes are in a remote wilderness provincial park, accessible only by aircraft, boat, or canoe. Photograph by D. Smith, ISPG 4236-5

about 50 km². The sand-dune activity here is unique, as it occurs within an area otherwise dominated by boreal forest. It is believed that dune activity was initiated about 6000 years ago as a result of winds reworking former lake deposits, and that some dunes have remained active ever since. A major factor in the activity of these dunes is the locally deep water table attributed to the great thickness of sand, which leads to comparatively dry conditions, thus permitting continued sand-dune activity in an otherwise humid region of Canada.

Other areas of widespread sand-dune activity include the western margins of the Great Sand Hills in Saskatchewan and the eastern edge of the Brandon Sand Hills in Manitoba (Fig. 10). Both areas have much higher levels of sand-dune activity than what would be expected under existing climatic conditions, and this may be the result of more arid conditions that existed in the past. Although the droughts of the 1930s and late 1980s seem to have led to relatively minor increases in dune activity in these areas, earlier droughts may have been so severe as to result in 'widespread' sand-dune activity. In general, these areas show a trend towards stabilization, with the area of active sand dunes decreasing over the last 50 to 70 years. However, sand-dune activity would likely increase again if the climate becomes drier again.

Coastal dunes

Coastal sand dunes in both lakeshore and maritime settings are an important part of many coastal ecosystems. Dunes are widespread on the Atlantic coast and along the Great Lakes, and to a lesser extent along other shorelines in Canada (Fig. 2). They commonly act to protect coastlines from wave action by creating an elevated shoreline. They typically absorb wave energy and release sediment back to the beach, thereby acting as important buffers against flooding and erosion of adjacent landward areas during storms. Dunes also provide diverse habitats for both plant and animal species. Despite this, coastal dunes are probably the most heavily affected dune areas in Canada. Logging, agriculture, sand mining, and intense recreational use have had negative impacts on coastal dunes (Fig. 11). In the 1950s and 1960s, numerous parks were established to meet growing recreational demands on beaches around the Great Lakes and in Eastern Canada. Little attention was paid to the sensitivity of coastal dunes at the time, and many dunes were damaged by overtrampling, unrestricted camping, and poorly located facilities. The resulting erosion of coastal sand dunes and the disruption of coastal ecosystems were difficult and costly to repair. In more recent years, because of increased concern for natural areas, protection strategies including dune management, designated wilderness areas, and information

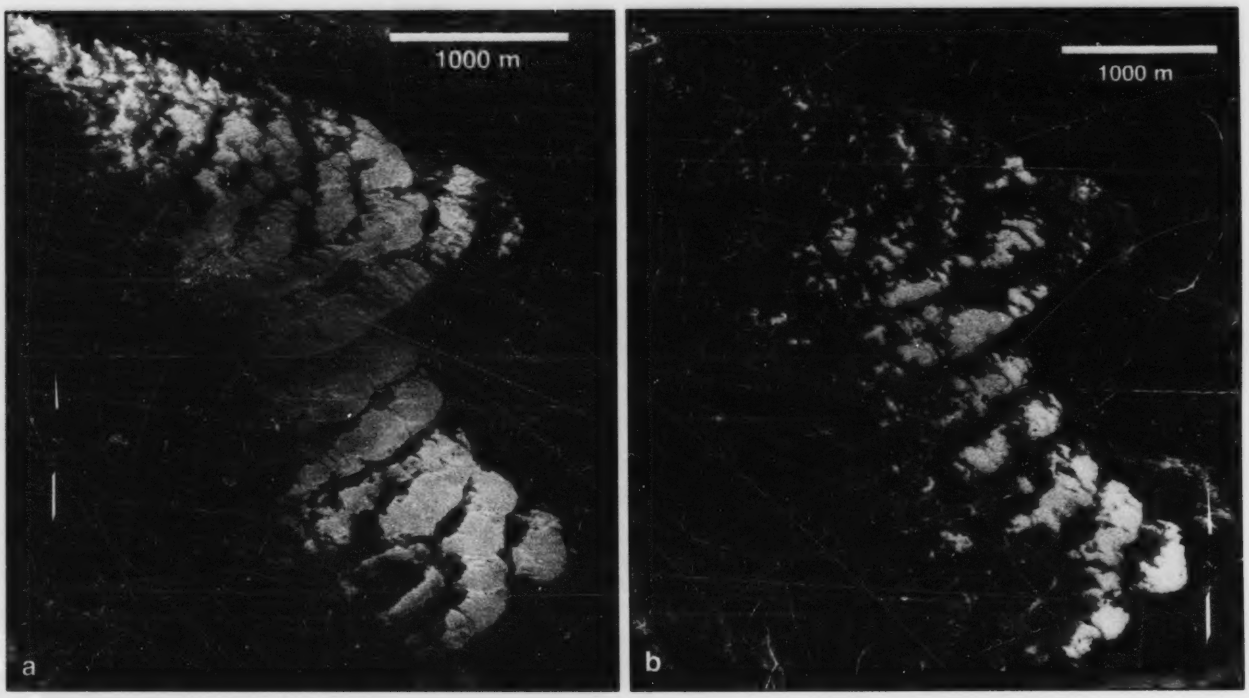


*Figure 10. 'The Devil's Hills' in Spruce Wood's Provincial Park, southeast of Brandon Manitoba, **a**) in 1954 (NAPL A12983-92) and **b**) in 1991 (NAPL A27716-137). The largest area of active sand dunes in Manitoba is in Spruce Woods Provincial Park and on the Shilo Military Reserve southeast of Brandon. In recent years, these dunes have been stabilizing at a rapid rate, with the area of active dunes decreasing by over 80% in the last 70 years. This stabilization suggests that the dunes probably owe their existence to more arid conditions in the past, and that they could reactivate if these conditions should recur.*

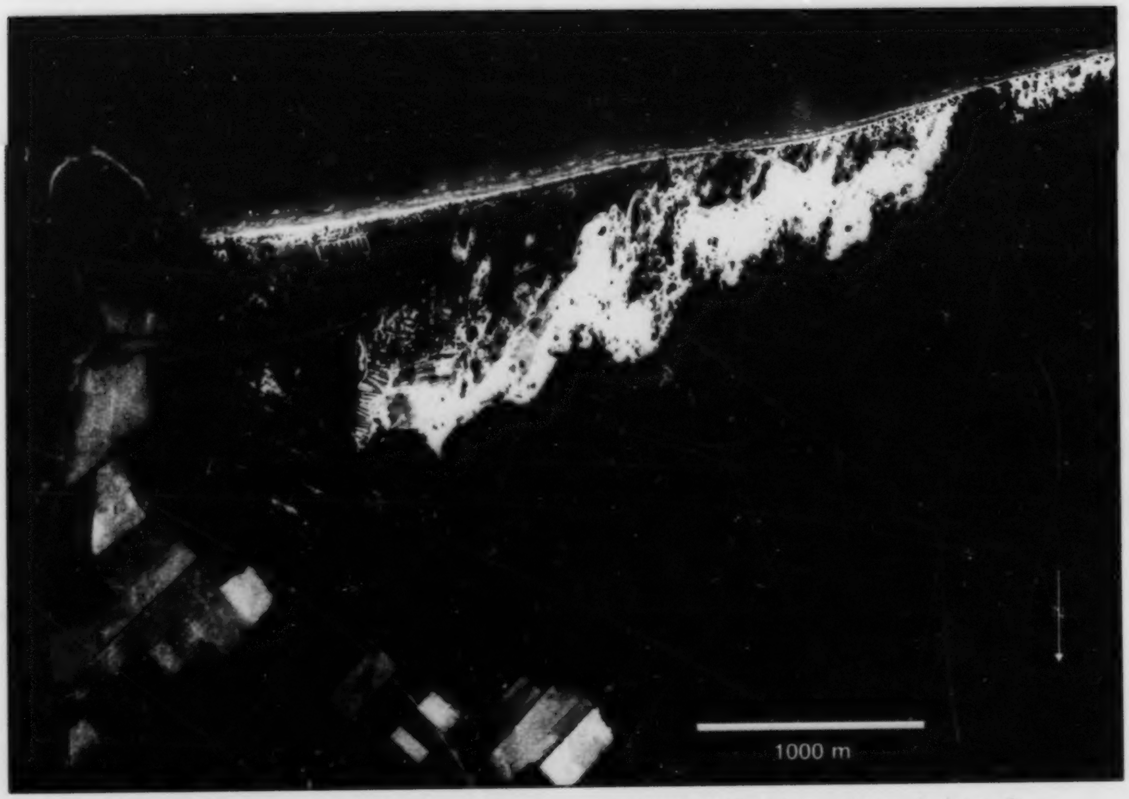


Figure 11. Coastal sand dunes at Sandbanks Provincial Park, Ontario, on the north shore of lake Ontario, have formed across a coastal spit extending from the shore. They protect the bay area from storm waves and erosion and represent an important part of the coastal ecosystem. They are also a popular recreational area and are thus heavily affected, especially in summer. Consequently, protection and management strategies have been instigated in this and other parks to reduce disturbance of the dunes. NAPL 21188-137

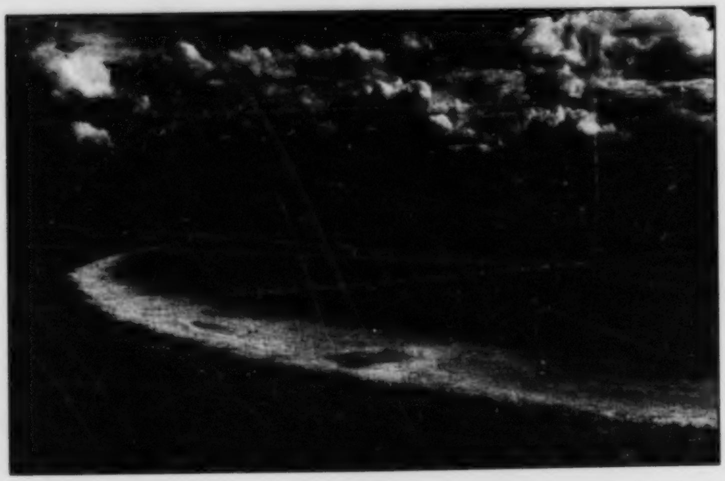


Figure 12.

One of the many of alkali lakes occurring on the Canadian Prairies, south of Golden Prairie, Saskatchewan, July 1994. Southern Saskatchewan has many small lakes such as this one, most of which are alkaline (contain salt). Recurrent droughts combined with water diversion contributes to the drying out of these shallow lakes, leaving only salt-covered basins. Dust-sized salt particles containing sodium, magnesium, sulphate and other noxious elements may be blown from these basins, creating health problems for humans and cattle nearby. Although the hazard is often short lived in Canada, it is a significant problem in some arid regions of the United States. GSC 1998-009D awareness programs have become an integral part of most park programs. Nevertheless, as these areas continue to be popular, parks will be under constant pressure to provide a balance between recreation and conservation.

Deflation plains

Deflation plains are areas eroded by wind, but devoid of sand dunes or sand sheets. Active deflation plains are restricted to High Arctic areas of strong winds, sparse vegetation cover, and little available sediment (Fig. 2). Very little is known about these areas, but sediment supply appears to be so limited that sand dunes do not form. Strong winds, blowing sand, and dust can reduce visibility on the ground and in the air. Currently, these conditions pose only a minor hazard to humans, as they occur in the more remote regions of Canada. With continued exploration, development, and habitation of Arctic regions, however, the potential impact of eolian activity will need to be considered.

Alkali dust emissions

Recurrent droughts in the 1980s, combined with ongoing water diversion, have contributed to the drying of many shallow lakes on the southern Prairies (Fig. 12). For example, Old Wives Lake in southern Saskatchewan, once the fourth largest alkaline lake in North America, was a 180 km² dry lake bed in the late 1980s. Dust containing high concentrations of sodium, magnesium, sulphate, chloride, phosphorus, and ammonium was blown from the dry lake bed across nearby prairie pastures and farmland, creating health problems for cattle and humans downwind. Exposure to the airborne salts caused weight loss and nasal and eye irritations in cattle, and nasal, throat, respiratory, and eye problems in people. Although this phenomanon was short lived, similar problems occur near dry lakes in drier regions of the United States and the former Soviet Union and result in chronic health problems for people living in these areas. Although the occurrence of alkali dust emissions is not shown on Figure 2, the problem is restricted mainly to the southern Prairies.

The future

Although wind erosion is now under control in most regions of Canada, projections of future climate warming and increased aridity in some areas suggest that the risk of wind erosion and sand-dune activity could increase in the future. The risk is probably greatest on the Canadian Prairies, where climate models predict that mean annual temperatures could increase by as much as 4°C, with soil-moisture availability remaining the same or decreasing slightly. Over the last 100 years, there has been a statistically significant increase in mean annual temperature in Western Canada with no discernable trend in precipitation. No clear evidence currently exists of increased wind erosion or sand-dune activity in Canada. Nevertheless, modelling suggests that increased aridity due to prolonged drought or climatic warming will lead to increased dune activity on the Canadian Prairies. In the driest regions, where dune activity is now greatest, this increased dune activity would likely result in an increased number of sand-dune blowouts, in reactivation of stabilized dunes, and in the merging of currently active sand dunes to form a nearly continuous bare sand cover in some areas. Climate change, however, is unlikely to result in desert-like conditions anywhere in Canada, although currently stabilized and partly active sand-dune areas will likely be more susceptible to human and natural disturbances. To counter this, diligent management practices including modified tillage and cropping practices and reduced grazing densities would likely be necessary on agricultural soils and grazing lands prone to wind erosion.

ACKNOWLEDGMENTS

The author is indebted to Jan Aylsworth, Cam Baker, Peter David, Lynda Dredge, Art Dyke, Louise Filion, Doug Hodgson, Rod Klassen, Lionel Jackson, Victor Levson, Beth McClenaghan, Yves Michaud, John Shaw, Robert Taylor, and Elaine Wheaton for providing reviews and contributions to the map of eolian activity in Canada. This contribution benefited from comprehensive reviews by Don Lemmen and Greg Brooks.

REFERENCES

Sparrow, H.O. (chairman)
1984: Soil at risk, Canada's eroding future; Standing Senate Committee on Agriculture, Fisheries and Forestry, Committees and Private Legislation Branch, The Senate of Canada, Ottawa, 114 p.
Wolfe, S.A. and Nickling, W.G.
1997: Sensitivity of eolian processes to climatic change in Canada; Geological Survey of Canada, Bulletin 421, 30 p.

SUGGESTED READING

Boer, G.J., McFarlane, N.A. and Lazare, M.
1992: Greenhouse gas-induced climate change simulated with the CCC second-generation General Circulation Model; Journal of Climate, v. 5, p. 1045—1077.
Coote, D.R. and Padbury, G.A.
1987: Saskatchewan preliminary wind erosion risk map; Land Resource Research Centre, Research Branch, Agriculture Canada, scale 1 000 000.
Coote, D.R., Padbury, G.A., Ellers, R.G., and Langman, M.N.
1987a: Manitoba wind erosion risk map; Land Resource Research Centre, Research Branch, Agriculture Canada, scale 1 000 000.
Coote, D.R., Padbury, G.A., and Pettapiece, W.W.
1987b: Alberta wind erosion risk map; Land Resource Research Centre, Research Branch, Agriculture Canada, scale 1 000 000.
Davidson, R.J.
1990: Protecting and managing Great Lakes costal dunes in Ontario; *in* Proceedings, Canadian Symposium on Coastal Sand Dunes, Guelph, Ontario, National Research Council, p. 455–471.
Environment Canada
1993: Canadian climate normals 1961–1990, Prairie Provinces; Minister of Supply and Services, Ottawa, 266 p.
Gullet, D.W. and Skinner, W.R.
1992: The state of Canada's climate: temperature change in Canada 1895–1991; Atmospheric Environment Service, Environment Canada, State of the Environment Report no. 92-2, 36 p.
1995: The state of Canada's climate: monitoring variability and change; Atmospheric Environment Service, Environment Canada, State of the Environment Report no. 95-1, 52 p.

Hind, H.Y.
1860: Narrative of the Canadian Red River exploring expedition of 1857 and of the Assinniboine and Saskatchewan exploring expedition of 1858 (two volumes); Longman, Green, Longman and Roberts, London, England, 425 p.

Schlesinger, M.E.
1987: Model projections of the equilibrium and transient climatic changes induced by increased atmospheric CO_2; *in* Proceedings, Symposium/Workshop on The Impact of Climatic Variability and Change on the Canadian Prairies, September 9–11, Alberta Research Council, p. 163–185.

Turner, A.H.
1955: How Saskatchewan dealt with her "Dust-Bowl"; Geographical Magazine, v. 28, p. 182–192.

Wheaton, E.E.
1992: Prairie dust storms — a neglected hazard; Natural Hazards, v. 5, p. 53–63.

Wheaton, E.E. and Chakravarti, A.K.
1990: Dust storms in the Canadian Prairies; International Journal of Climatology, v. 10, p. 829–837.

Wolfe, S.A.
1997: Impact of increased aridity on sand dune activity in the Canadian Prairies; Journal of Arid Environments, v. 36, p. 421–432.

Wolfe, S.A., Huntley, D.J., and Ollerhead, J.
1995: Recent and late Holocene sand dune activity in southwestern Saskatchewan; *in* Current Research 1995-B; Geological Survey of Canada, p. 131–140.

Additional material used for map compilation

Chapman, L.J. and Putnam, D.F.
1973: The Physiography of Southern Ontario (second edition); University of Toronto Press, Toronto, Ontario, 386 p.

David, P.P.
1977: Sand dune occurrences in Canada: a theme and resource inventory study of eolian landforms in Canada; Indian and Northern Affairs, National Parks Branch, Contract no. 74-230,183 p.
1989: Eolian Processes; *in* Chapter 9 of Quaternary Geology of Canada and Greenland, (ed.) R.J. Fulton; Geological Survey of Canada, Geology of Canada no. 1, p. 620–623 (*also* Geological Society of America, The Geology of North America, v. K-1, p. 620–623).

Filion, L. and Morisset, P.
1983: Eolian landforms along the Eastern coast of Hudson Bay, Northern Quebec; *in* Tree-Line Ecology, (ed.) P. Morisset and S. Payette; Proceedings, Northern Quebec Tree-Line Conference; Collection Nordicana no. 47, p. 73–94.

Hales, W.J.
1993: Coastal sand dunes of Nova Scotia; M.Sc. thesis, McMaster University, Hamilton, Ontario, 271 p.

Martini, I.P.
1981: Coastal dunes of Ontario: distribution and geomorphology; Géographie physique et Quaternaire, v. 35, p. 219–229.

McCann, S.B.
1990: An introduction to the coastal dunes of Atlantic Canada; *in* Proceedings, Symposium on Coastal Sand Dunes, Guelph, Ontario, National Research Council, p. 89–107.

Pissart, A., Vincent, J.-S., and Edlund, S.A.
1977: Dépôts et phénomènes éoliens sur l'île de Banks, Territoires du Nord-Ouest, Canada; Journal canadien des sciences de la terre, vol. 14, p. 2462–2480.

Permafrost in Canada, a challenge to northern development

S.L. Smith[1], M.M. Burgess[1], and J.A. Heginbottom[1]

Smith, S.L., Burgess, M.M., and Heginbottom, J.A., 2001: Permafrost in Canada, a challenge to northern development; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 241–264.*

Abstract: The permafrost zone covers about half of the Canadian landmass and presents unique challenges to development in the north. Within this zone, permafrost continuously underlies the land in the far north and may be several hundred metres thick. At the southern limit of the zone, however, permafrost distribution is discontinuous and thickness is generally less than 10 m. Ground stability problems associated with thawing of ice-rich permafrost can occur in response to disturbance of the ground surface or climate change. Current knowledge about the distribution and characteristics of permafrost, including its thickness and temperature regime, and of ground ice in Canada is summarized. This information is important for rational planning of development in the north. Numerous examples, including transportation systems, buildings and foundations, utilities, pipelines, dams, and dykes, illustrate the challenges to development encountered in the permafrost region and the design and construction strategies used. The potential effects of climate change on permafrost are discussed briefly.

Résumé : La zone à pergélisol couvre environ la moitié de la masse continentale du Canada et présente des défis bien particuliers au développement du Nord canadien. Le pergélisol est continu dans le Grand Nord où il peut atteindre plusieurs centaines de mètres d'épaisseur. Cependant, à l'extrémité sud de la zone, le pergélisol est discontinu et son épaisseur est généralement inférieure à 10 m. Des problèmes de stabilité du sol associés au dégel du pergélisol riche en glace peuvent se produire en réponse aux perturbations de la surface du sol ou au changement climatique. L'article présente un aperçu des connaissances actuelles sur la répartition et les caractéristiques du pergélisol, notamment son épaisseur et son régime de température, et sur la glace de sol au Canada. Ces données sont importantes pour la planification rationnelle du développement du Nord canadien. De nombreux exemples, notamment les systèmes de transport, les édifices et les fondations, les services publics, les pipelines, les barrages et les digues, illustrent les défis auxquels est confronté le développement des régions pergélisolées et permettent de présenter les stratégies de construction et de conception appliquées dans ce contexte. Les incidences potentielles du changement climatique sur le pergélisol sont également brièvement présentées.

---

[1] Terrain Sciences Division, Geological Survey of Canada, Ottawa, Ontario K1A 0E8

INTRODUCTION

The permafrost region covers 5 700 000 km² or about half the land area of Canada. Within this region, the presence of perennially frozen ground and associated ground ice strongly influences the properties and performance of ground materials, whether soil or rock, and also controls to a significant degree the development of the land.

The distribution, thickness, and stability of permafrost is controlled by climate, ground temperature, and the nature of ground materials. For nearly all forms of development in northern regions, the occurrence of ground ice within the permafrost is a prime concern. This is because of ground-stability problems associated with the thawing of ice-rich permafrost as a result of disturbance of the ground surface or climate change. Ground-ice formation associated with permafrost growth or with seasonal freezing can also present problems, such as frost heave, that need to be considered when planning development. Accurate information on the character, distribution and form of frozen ground and ground ice, the geographical and geological setting, climatic conditions, properties of subsurface materials and the geothermal environment are all important for the rational planning of development in northern Canada.

This contribution summarizes current knowledge about the distribution and characteristics of permafrost and ground ice in Canada as well as permafrost thickness and ground temperature. The constraints posed by permafrost on development are also discussed. Numerous examples are presented to illustrate the challenges to development that are encountered in the permafrost region and the design and mitigation strategies used. These examples focus on transportation systems (roads, railways), buildings and foundations, utilities, pipelines, dams, and dykes. The discussion is not exhaustive. For example, issues relevant to onshore and offshore hydrocarbon exploitation, offshore pipelines, and mineral exploitation are not included. Hazards associated with natural gas hydrates in permafrost environments are outlined in the contribution 'Natural gas hydrates'. Finally, a brief discussion of permafrost and climate change is provided. Figure 1 shows the location of communities and mines mentioned in this contribution.

PERMAFROST DISTRIBUTION AND GROUND THERMAL REGIME

Terminology and definitions used in this commentary generally follow those accepted by the International Permafrost Association (1998). The fundamental science behind the

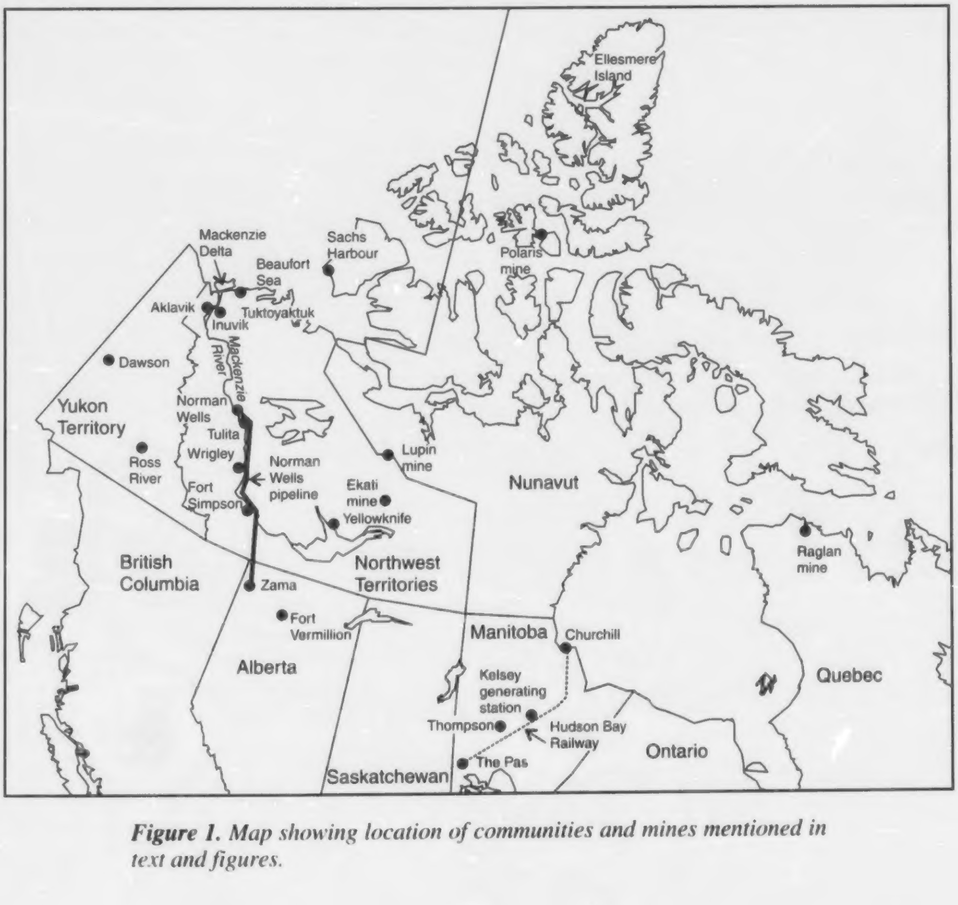

Ellesmere Island
Sachs Harbour
Polaris mine
Mackenzie Delta
Beaufort Sea
Aklavik
Inuvik
Tuktoyaktuk
Dawson
Mackenzie River
Yukon Territory
Norman Wells
Tulita
Wrigley
Ross River
Fort Simpson
Norman Wells pipeline
Lupin mine
Ekati mine
Yellowknife
Nunavut
Raglan mine
British Columbia
Zama
Northwest Territories
Fort Vermillion
Manitoba
Churchill
Alberta
Kelsey generating station
Quebec
Thompson
Hudson Bay Railway
Ontario
The Pas
Saskatchewan

Figure 1. Map showing location of communities and mines mentioned in text and figures.

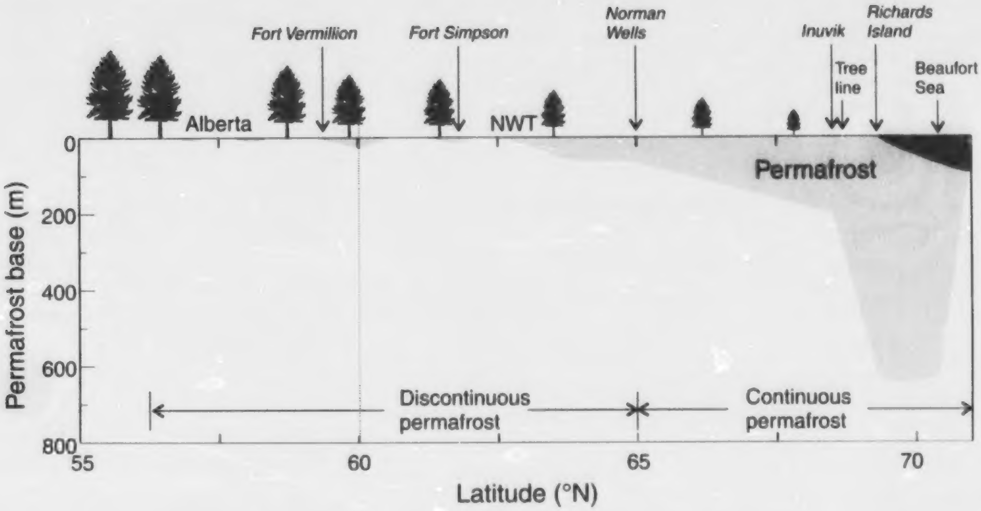


Figure 3. Latitudinal variation in permafrost thickness from northern Alberta through the Mackenzie River valley to the offshore region of the Beaufort Sea. The Richards Island area and adjacent Beaufort Shelf were exposed to cold temperatures during the Wisconsinan Glaciation due to the absence of glacial ice or recession of seawater. This allowed very thick permafrost to develop in these areas. In areas to the south and west, ice sheets insulated the ground from the subzero temperatures of the Wisconsinan Glaciation and thick permafrost could not develop.

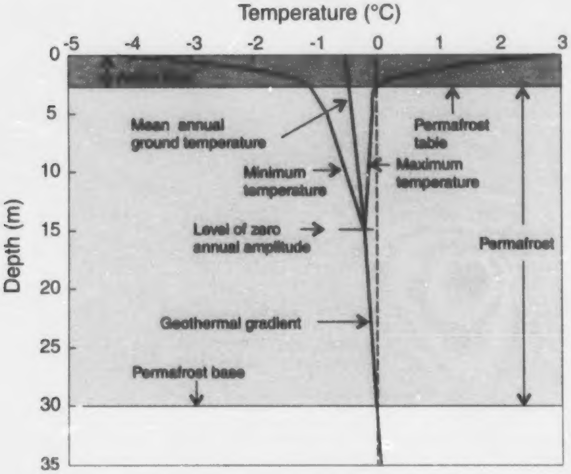


***Figure 4.** Features of the ground thermal regime in permafrost based on data from a Yellowknife site. The annual ground-temperature envelope which is defined by the maximum and minimum temperatures at each depth, is shown. Also shown is the mean annual ground-temperature profile.*

various concepts is well reported in several textbooks, examples of which are included in the 'Suggested reading' list at the end of this contribution.

Permafrost is defined as ground, whether soil or rock, that remains at or below 0°C for at least two consecutive years. Most permafrost has existed for much longer and in the far north, thicknesses reach several hundred metres.

At any location, permafrost thickness depends on the temperature at the ground surface, the geothermal heat flow from the Earth's interior, which is a function of the geological environment and history, and the ground thermal properties. Permafrost thickness ranges from a few metres or less at the southern limit of permafrost to over 700 m in the Arctic Islands (Fig. 2a in pocket, 3).

The ground-temperature regime in a region with permafrost is illustrated in Figure 4. The annual ground-temperature envelope is defined by the maximum and minimum temperatures reached at each depth during the year. Daily air-temperature changes penetrate only a few tens of centimetres below the ground surface. At depths of a few metres, only the annual change in air temperature from summer to winter is observed. The seasonal variation in ground temperature may be quite large at the ground surface, but decreases with depth and is completely dampened out at depths of approximately 10 to 15 m. The level of zero annual amplitude is the depth at which there is no observed seasonal variation in ground temperature. The natural damping effect of the ground is enhanced by snow, water, vegetation, and organic material, which tend to further buffer the ground temperature from changes in air temperature. Fluctuations in air temperature generally penetrate to shallower depths when a significant buffer layer is present between the air and the ground. Below the level of zero annual amplitude, temperature increases with depth according to the geothermal gradient. The mean annual ground-temperature profile is defined by the average temperature at each depth. The mean annual temperature of permafrost in the upper 10 m of the ground ranges from 0°C to approximately -20°C over the range of latitude of 57°N to 83°N (Fig. 5).

The upper part of the ground that thaws each summer and refreezes each winter constitutes the active layer, which technically is not part of the permafrost. Climate and local factors such as snow cover, vegetation, presence or absence of an organic layer, soil moisture, and thermal properties of surficial materials determine the thickness of the active layer. In areas of exposed bedrock, for example, the active layer may be several metres thick. In vegetated, organic terrain, under similar climatic conditions, the active layer may be less than 0.5 m thick.

The atmospheric climate is the main factor determining the occurrence of permafrost. The relationship between the atmospheric climate and the ground thermal regime depends on site-specific conditions that determine the transfer of heat between the ground and the air. These conditions include vegetation, snow cover, surficial materials, the presence or absence of an organic layer, soil moisture content, and drainage. They determine the ground-surface temperature on which the spatial distribution, thickness, and temperature of permafrost are highly dependent. The mean annual near-surface ground temperature for over 300 sites has been extracted from a Geological Survey of Canada (GSC) database and is shown in Figure 2b. Although the spatial pattern of ground and air temperature is similar, mean annual ground-surface temperature is generally 2 to 4°C higher than mean annual air temperature (Fig. 5, 6) and exhibits much greater local variation.

As noted, permafrost underlies about half of Canada's landmass and areas of the seabed in the Western Arctic. It is also believed to exist beneath the channels of the Arctic Islands. Conditions for permafrost existence prevail not only at high latitudes, but also at high altitudes, mainly in the mountains of Western Canada.

Across the arctic regions of Canada, permafrost underlies essentially all exposed land areas and is therefore termed 'continuous permafrost' (*see* Fig. 2c, *in* pocket). In this zone of continuous permafrost, unfrozen ground is found only beneath the larger lakes and rivers and in small areas of newly aggraded ground. Bodies of unfrozen ground above, within, or through permafrost are termed 'taliks'. Taliks may be found, for example, beneath small water bodies. South of this zone lies a broad zone of discontinuous permafrost, in which the proportion of unfrozen ground increases southward (Fig. 2c). This zone is commonly subdivided into a zone of extensive discontinuous permafrost, wherein over 50 % of the land area is underlain by permafrost, and a zone of sporadic discontinuous permafrost, wherein 10 to 50% of the land area is underlain by permafrost. Farther south is a zone in which permafrost is confined to isolated patches of terrain, in locally favourable locations such as areas of elevated organic terrain (Fig. 7).

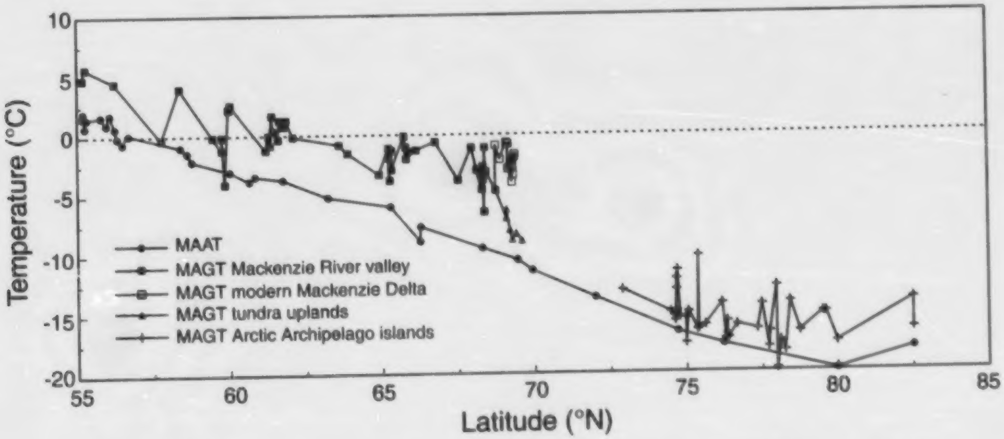


Figure 5. *Latitudinal variation in mean annual air temperature (MAAT) and near-surface ground temperature (MAGT) from northern Alberta through the Mackenzie River valley to the Mackenzie Delta. Ground temperatures in the Mackenzie Delta region are shown for the modern delta and the adjacent tundra uplands. Air and ground temperatures are also shown for the islands of the Arctic Archipelago. Air temperature data are from the Atmospheric Environment Service climate normals and ground temperature data are from a GSC database. In the Mackenzie Delta area, ground temperatures are considerably warmer on the modern delta than in the adjacent tundra uplands. Vegetation, which leads to deeper snow cover, and the large amount of water cover in numerous lakes and river channels are the main factors contributing to the warmer ground temperatures on the modern delta. Note: points are joined with lines to highlight differences between each location; the lines are not necessarily representative of interpolated temperatures.*

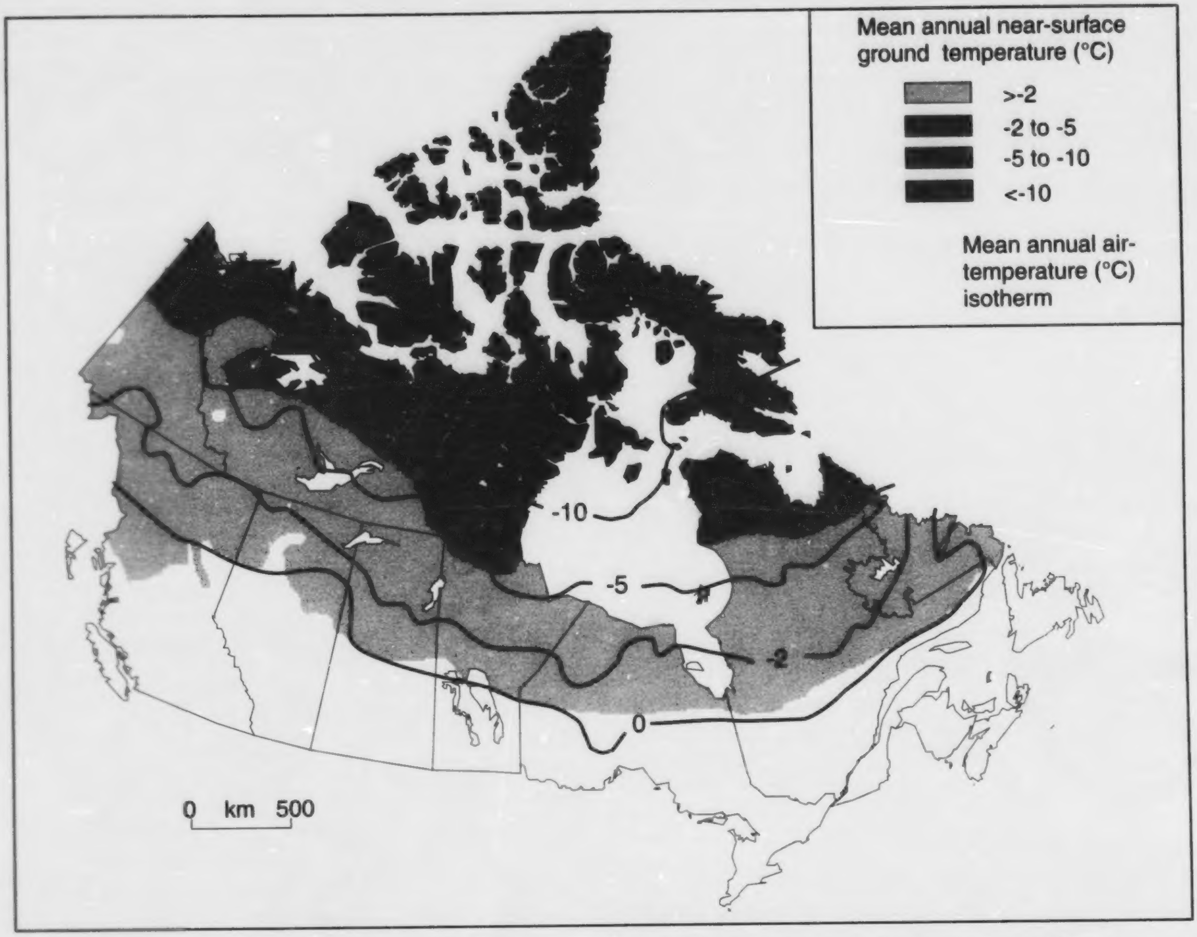

Mean annual near-surface ground temperature (°C)
>-2
-2 to -5
-5 to -10
<-10
Mean annual air-temperature (°C) isotherm
-10
-5
-2
0
0 km 500

Figure 6. *Spatial variation in mean annual air temperature and near-surface ground temperature within the Canadian permafrost region. Air temperature data are from Atmospheric Environment Service (1982). Ground temperature within the zone classified as over -2°C will only be colder than 0°C where permafrost is present. This zone is within the discontinuous permafrost zone.*

Figure 7.

Norman Wells pipeline right-of-way crossing organic terrain in the discontinuous permafrost zone of northern Alberta. The alternation of light (treeless and unfrozen) and dark (forested and frozen peat plateau) segments illustrates the discontinuous nature of the permafrost in this area. Photograph by M.M. Burgess, July 1988. GSC 2000-166

The body of subsea permafrost beneath the Beaufort Sea (Fig. 2c) is a relict of periods of lower sea level during the glacial maxima of the Quaternary Period. During these periods, large areas of the continental shelf were above sea level and exposed to the intense cold of a full glacial climate. During the interglacial stages and in postglacial time, these areas were covered by Arctic Ocean water with a mean annual temperature 10 to 15°C higher than air temperatures during glacial periods. The thermal regime of the subsea permafrost is thus in disequilibrium with the current marine environment. The sediments therefore are warming gradually, causing the permafrost to slowly degrade.

GROUND ICE IN PERMAFROST REGIONS

At temperatures below 0°C, almost all moisture in the ground occurs in the form of ice. Compared to most other earth materials, ground ice exists at temperatures close to its melting point. In permafrost regions, ground ice is one of the most important features of the terrain. Its presence influences topography, geomorphic processes, vegetation, and landscape response to environmental changes, whether natural or anthropogenic. A key concept in understanding the influence of ground-ice conditions in permafrost areas is that of excess ice, or ice whose volume, when melted, exceeds the void volume or pore volume of the enclosing sediments. Freezing and thawing processes and the presence of excess ice are responsible for the many landscape features associated with ground ice. Ground ice occurs as structure-forming ice, bonding the enclosing sediment, and as larger bodies of more or less pure ice. Structure-forming ice includes pore ice, icy coatings on soil particles, ice veins, ice lenses, and intrusive ice. Large bodies of more or less pure ice occur as ice wedges, as ice in ice-cored mounds and pingos, and as massive ice (Fig. 2d, e, f, 8).

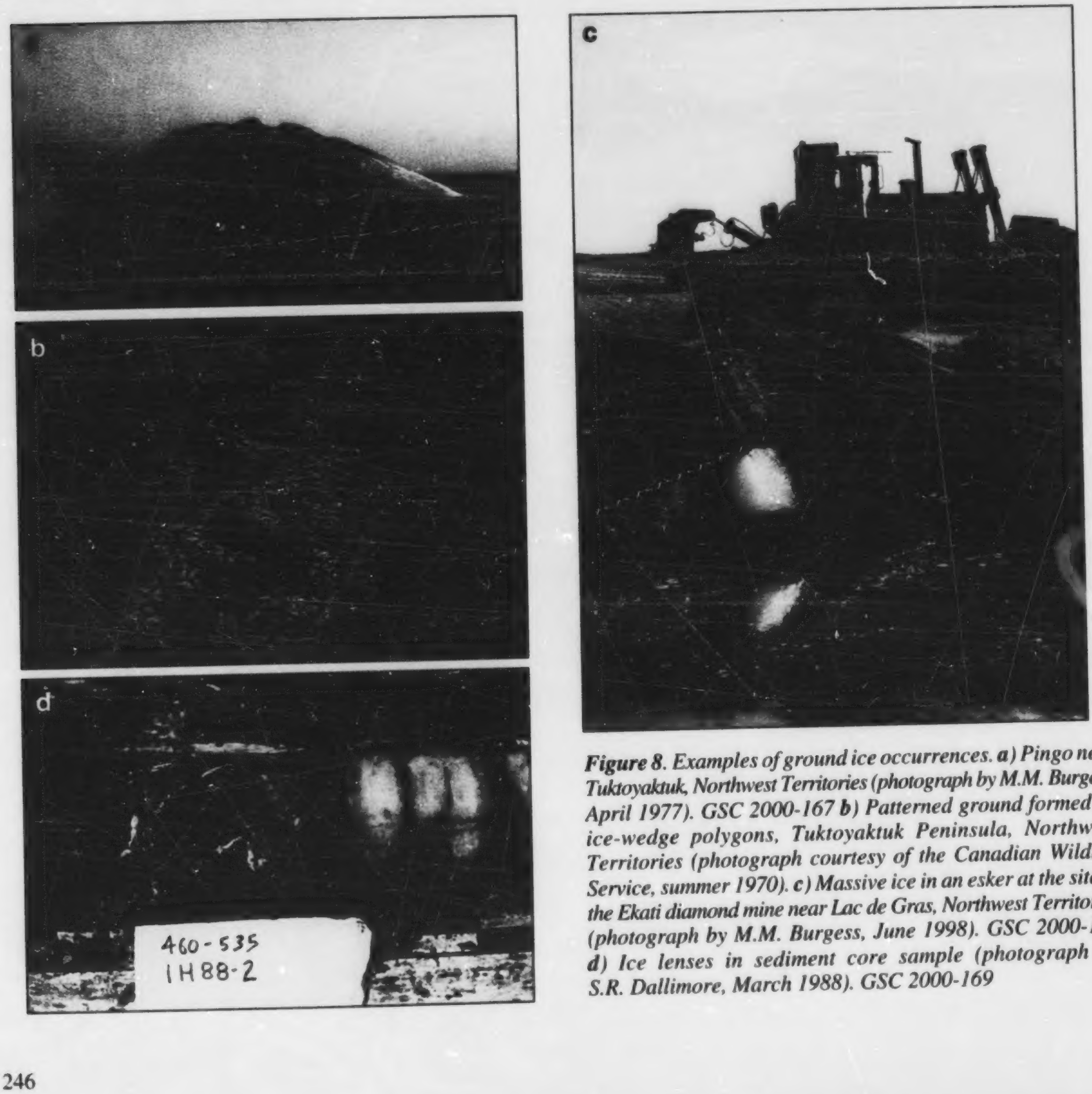


Figure 8. Examples of ground ice occurrences. ***a)*** *Pingo near Tuktoyaktuk, Northwest Territories (photograph by M.M. Burgess, April 1977). GSC 2000-167* ***b)*** *Patterned ground formed by ice-wedge polygons, Tuktoyaktuk Peninsula, Northwest Territories (photograph courtesy of the Canadian Wildlife Service, summer 1970).* ***c)*** *Massive ice in an esker at the site of the Ekati diamond mine near Lac de Gras, Northwest Territories (photograph by M.M. Burgess, June 1998). GSC 2000-168* ***d)*** *Ice lenses in sediment core sample (photograph by S.R. Dallimore, March 1988). GSC 2000-169*

The quantity of ice in the ground varies widely. At one extreme, it can make up 50 to 70% by volume of the upper 2 to 3 m of permafrost over broad areas; locally, it can exceed 90% of the volume of the ground. In other areas, permafrost contains essentially no ground ice and is termed 'dry'. The distribution of ground ice is strongly influenced by soil texture and geological and geothermal history. Generally, fine-grained, clay- and silt-rich soils and organic soils contain more ground ice than coarse-grained soils consisting of sand and gravel. The most ice-rich permafrost tends to be found within the upper layers of the ground, which is often the area of greatest concern for engineering and construction. The major categories of ground ice are described below and their distribution is shown in Figure 2c, d, e, and f.

Pore ice

Pore ice includes individual crystals of ice within the pores of soils and rocks, and icy coatings on soil particles. The amount of pore ice in a soil depends on soil porosity, which in fine-grained silts, for example, may be about 40% of the total soil volume. On melting, pore ice does not yield more water than the void volume of the same soils or rock when unfrozen. Permafrost is said to be 'ice bonded' if sufficient pore ice is present to act as cement and bond the soil particles together.

Ice lenses and veins

Most ground ice occurs in the form of ice lenses, which are dominantly horizontal, thin sheets with tapering edges. Ice lenses may vary in thickness from a hairline to over 10 m, but very thick and extensive ice lenses are better termed 'massive-ice beds'. The term 'ice lens' is commonly used for a layer of segregated ice that is more or less parallel to the ground surface. Ice lenses are commonly located at the interfaces of changes in soil texture and tend to develop preferentially at the base of finer layers and above coarse layers. Ice veins normally cut across the soil stratigraphy and develop in tension or dilation cracks, or as a result of water intrusion. In clay soils, the freezing process may produce reticulate vein ice, which is a network of veins of relatively pure ice surrounding soil blocks that contain little ice. It is generally assumed that most vein or lens ice is formed in place by segregation (*see* below) or injection.

Ice wedges and ice-wedge polygons

Freezing and thawing of the active layer, coupled with temperature changes in the upper part of the permafrost, result in the development of a suite of distinctive landforms, collectively known as 'patterned ground'. These include primarily ice-wedge or frost polygons.

Ice wedges develop from the repeated incursion of snow or water into cracks in the ground that form as a result of thermal contraction in winter. The resulting ice persists and creates a zone of weakness, which localizes the point of cracking in subsequent years. Over decades, the ice wedge widens and ridges of displaced soil develop on either side of a trough, giving a distinctive microtopography. The spatial pattern of cracking results in a network of ice wedges and a field of ice-wedge polygons.

Pingos and frost mounds

A pingo is a large, perennial frost mound with a core of massive ice, produced primarily by injection of water and covered with soil and vegetation. Pingos are conical or hemispherical and range from a few metres to as much as 50 m in height and from a few tens of metres to several hundred metres in basal diameter or length. The top is often cratered or cracked due to rupturing of the soil and vegetation cover as the ice core grows.

Closed-system pingos are found in flat, poorly drained terrain of the continuous permafrost zone. They form by uplift of the ground surface due to freezing of injected water supplied by the expulsion of pore water during permafrost aggradation in a talik at the site of a drained lake or pond. Over 1350 closed-system pingos are found in the Tuktoyaktuk coastlands of northwestern Canada.

Open-system pingos are generally found in areas of high local relief, mainly in the discontinuous permafrost zone. They form by uplift of the ground due to freezing of injected water supplied by groundwater moving downslope through taliks to the pingo site. In Canada, open-system pingos are most common in west-central Yukon Territory, where about 400 have been reported.

Smaller forms of frost mounds include palsas and frost blisters. A palsa is a peaty permafrost mound with a core of layers of segregated ice and peat or soil. A frost blister is a seasonal frost mound with a core of ice that forms under conditions of high hydraulic pressure during freezing of the active layer.

Massive-ice beds

Thick and extensive ice lenses, termed 'massive-ice beds', can be up to tens of metres thick and hundreds of metres in horizontal extent. In Canada, such ice forms are widespread in the Arctic Lowlands and seem to be responsible for much of the visible local relief. Massive-ice beds are possibly formed by ice-segregation processes (*see* below) or by injection of water under pressure, or they may be buried deposits of surface ice such as glacier ice, river icings, snow banks, and lake ice.

DEVELOPMENT IN THE PERMAFROST REGION

Freezing and thawing of soils

The problems encountered during construction and development in permafrost regions are classified as 1) those involving the thawing of ice-rich materials under either unheated structures such as roads or heated structures such as buildings; 2) those resulting from frost action, such as frost heave;

and 3) those associated with the temperature of permafrost, for example the freezing of buried water lines. Since constraints to development and construction in the North are commonly associated with the melting of ground ice or freezing of soil water, it is important to understand what happens as soils freeze or thaw. A knowledge of the mechanical and physical properties of freezing and thawing soils is also required to properly design structures for cold regions.

As soil temperature falls below 0°C, water within the soil's pores freezes gradually over a range of temperatures below 0°C. Water and ice can therefore coexist in frozen soils, and appreciable amounts of unfrozen water may be present in fine-grained soils at temperatures down to several degrees below 0°C. Moisture is drawn from the warmer unfrozen soil towards the freezing soil. Ice lenses may develop within the frozen soil as this water accumulates and freezes. Lens ice is often referred to as 'segregated ice' because it is separated from the soil particles. As ice lenses grow, the ground surface is displaced upward, resulting in frost heave (Fig. 9). The frost heave process may cause the total water content of the frozen soil to increase beyond its unfrozen porosity, and the soil is then considered to contain excess ice.

The amount or rate of frost heave depends on soil water conditions, the rate of cooling of the soil, soil texture, and overburden pressure. Fine-grained, silty soils are generally the most susceptible to frost heave and are referred to as 'frost-susceptible soils'. Coarse-grained sand and gravel generally exhibit little or negligible heave and are termed 'frost-stable soils'.

Frost heave may occur during permafrost formation or during seasonal freezing. It may also occur beneath structures such as chilled pipelines. If conditions are uniform beneath the entire structure, then the movement is uniform and can often be tolerated by the structure. Soil and moisture conditions can, however, vary highly over relatively short distances, and differential movement commonly occurs that can result in damage to structures such as roads, airstrips, buildings, buried utility lines, and pipelines. Displacement due to frost heave is perpendicular to the freezing surface because ice lenses tend to form parallel to the freezing plane. Displacement can occur laterally, for example where a vertical face is exposed to freezing temperature, as would be the case in construction of a retaining wall.

Ice bonding is an important component of the strength of fine-grained frozen soils. As ground temperature increases, ice bonding decreases because of an increase in unfrozen water content, especially within a few degrees of 0°C. Thus, frozen soils weaken as they thaw and lose all strength due to ice bonding once they have thawed completely. The creep characteristics of ice and frozen soil are temperature dependent and need to be considered when loads are to be placed on frozen material. Ice and frozen soil will flow or deform slowly under sustained loads. Creep rates of ice-rich soils must be evaluated to determine whether they are within the design limits of a given structure.

Thawing of soil that contains excess ice in the form of ice lenses or bodies of massive ice can further weaken the soil because of an increase in pore-water pressures and a decrease

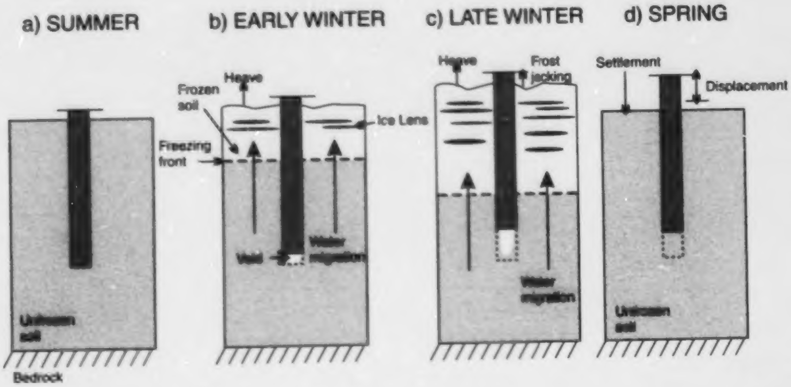


Figure 9. Four diagrams illustrate ice-lens formation and frost jacking of a pile in a freezing soil. ***a)*** *The original position of the pile after installation during the summer.* ***b)*** *As the soil freezes in early winter, water migrates from the unfrozen soil to the freezing soil. Because unfrozen water and ice coexist at temperatures below 0°C, water may travel a significant distance behind the freezing front before it accumulates and freezes in the form of ice lenses. Ice lens formation and expansion of the soil volume result in heaving of the ground surface. Adfreeze forces on the side of the pile cause the lifting or jacking of the pile, leaving a void that fills with soil below the base of the pile.* ***c)*** *Throughout the winter, the freezing front penetrates to greater depths and ice lenses continue to grow.* ***d)*** *Settlement of the ground surface and pile occurs as ground ice melts in the spring. The pile, however, cannot settle back to its original position and therefore undergoes a net upward displacement.* (Modified from *Wolfe, 1998*).

in strength if excess water is unable to drain away. As the water eventually drains away, thaw settlement will occur as the volume formerly occupied by the excess ice is lost (Fig. 10). Frozen soils that are subject to significant settlement during thaw are referred to as 'thaw unstable' or 'thaw sensitive'. Thaw-unstable soils include fine-grained material such as silt and clay and organic material or peat that generally have high ice contents and are also compressible in the unfrozen state. Differential thaw settlement of the ground can produce thermokarst topography, which is irregular terrain often containing of ponds or lakes.

The potential development hazard due to thaw settlement is greatest in regions of ice-rich permafrost. Information from the surficial geology and peatland maps of Canada has been used to produce a map (Fig. 11) characterizing thaw-settlement potential. A substantial part of the region containing warm permafrost (ground temperature above -2°C) has a high potential for thaw instability.

Slope stability

Warming of ice-rich soils and melting of ground ice on slopes can cause instability resulting in flows and slides. A flow is a landslide in which movement of debris has characteristics of a fluid. Flows include shallow active-layer detachments or skin flows, deeper retrogressive thaw flows, and rapid debris flows. Active-layer detachment flows are shallow slope failures involving the detachment and downslope movement of ground within the active layer and associated vegetation mat. They may be triggered by unusually warm temperatures or a disturbance to the vegetation cover, such as wildfire or clearing for construction (Fig. 12). Subsequent deepening of the active layer into ice-rich soil and impedance of excess moisture can result in detachment and movement downslope. Retrogressive thaw flows occur in ice-rich terrain and consist of a low-angle tongue of debris that extends downslope from a steep headwall. They may be initiated when massive ice or icy sediment is exposed to melting, as can occur following an active-layer detachment. Once the icy material is exposed in the headwall, thaw flows will occur each summer until the headwall face eventually becomes buried by debris and the slide stabilizes. Debris flows occur in areas of high relief and are a rapid, destructive flow of water-saturated sediment. They may be triggered by extreme precipitation events or unusually deep seasonal thaw in response to extremely warm summer temperatures.

Slides are landslide movements in which the slide components move downslope as a block or series of blocks (Fig. 13). The downslope movement of a block of sediment along a curving failure plane is called a 'rotational slide'. Undercutting of a river bank is a common trigger of rotational slides resulting in enough loading to rupture frozen sediments. As ice melts within the block, it may deform and flow.

Landslides in permafrost areas can damage pipelines, roads, bridges, and buildings; create landslide dams that interrupt river navigation or cause flooding; and increase siltation in streams which may affect fisheries. Such landslides can vary significantly in size; in the Mackenzie River valley, for example, they range from 220 m² to 1 800 000 m².

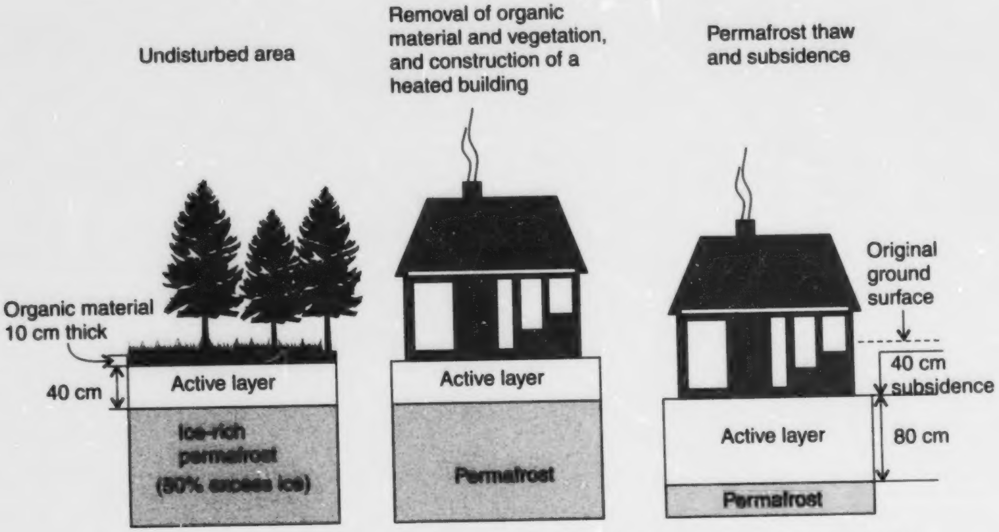

Undisturbed area
Removal of organic material and vegetation, and construction of a heated building
Permafrost thaw and subsidence
Organic material 10 cm thick
40 cm
Active layer
Ice-rich permafrost (50% excess ice)
Active layer
Permafrost
Original ground surface
40 cm subsidence
Active layer
80 cm
Permafrost

Figure 10. *Thaw settlement caused by disturbance of the ground surface and construction of a heated building. Prior to disturbance, the active layer was 50 cm thick. The thaw of ice-rich permafrost caused by the removal of vegetation and 10 cm of organic material and the construction of the building has increased active-layer thickness to 80 cm. The melting of ground ice and the drainage of water from the deepened active layer result in subsidence of the ground underneath the building. Note: diagram is not to scale* (modified from *French, 1996*)

Problems associated with development in the permafrost region

Ground instability is a major concern for development in permafrost regions, especially in areas where permafrost temperature is near 0°C. The removal of vegetation or insulating organic cover and other disturbances to the ground surface can result in changes to the ground thermal regime, leading to warming and melting of permafrost. In addition, permafrost can be warmed and melted by heat generated from heated buildings or buried water and sewage pipelines.

In areas of ice-rich, fine-grained sediment, failure to take proper precautions can have serious consequences for development. Following the discovery of gold in the Yukon

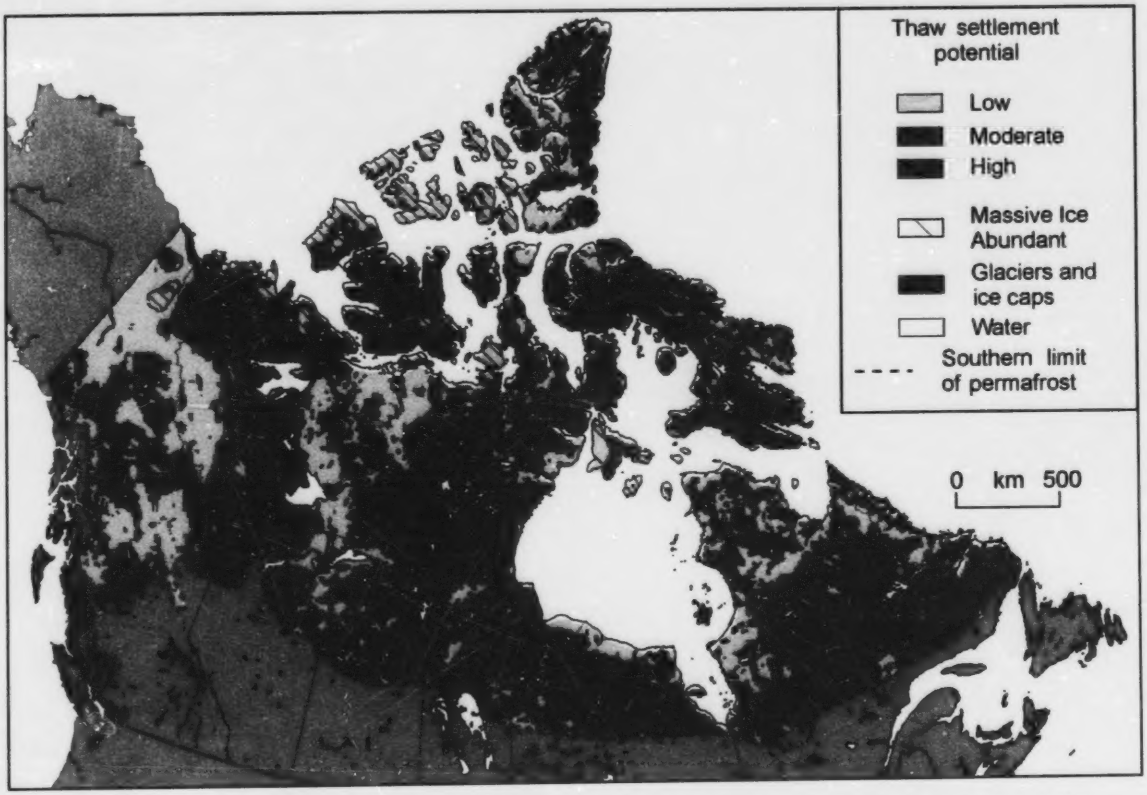


Figure 11. *Thaw-settlement potential determined by combining information on surficial materials and the presence of organic material. The thaw-settlement potential will be higher than suggested by the surficial material where massive ice (including wedge ice) is present.*

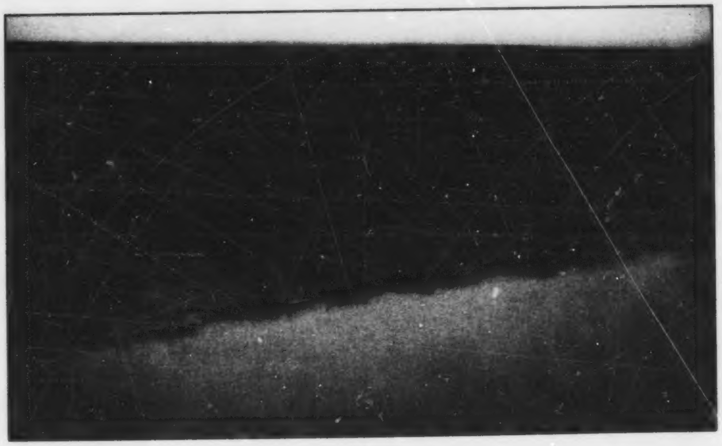


Figure 12.

Numerous active-layer detachments along a bank of the Mackenzie River between Norman Wells and Tulita, Northwest Territories. The area was burned by a forest fire in June 1995. Photograph by M.M. Burgess, September 1996. GSC 2000-034C

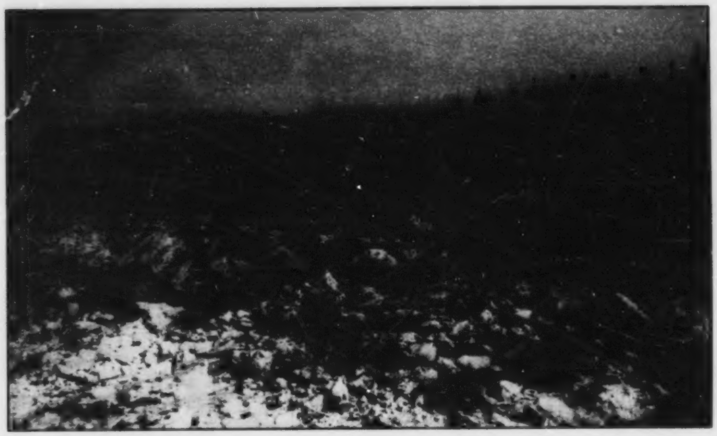


Figure 13.

Block landslide in permafrost along the Mackenzie River, near Old Fort Point Northwest Territories. The landslide occurred in the winter of 1997 and flowed onto the frozen river. Photograph by M.M. Burgess, March 1997. GSC 2000-170

Territory in the 1890s, construction in Dawson proceeded with no appreciation for the ice-rich alluvial silts underlying the site. Buildings were similar in design and construction to those in southern towns. The subsequent melting of ground ice and settlement of the underlying ground damaged buildings, many of which became uninhabitable. Roads became impassable because of differential settlement as underlying permafrost thawed. During the establishment of Aklavik, in the Northwest Territories, inappropriate southern building designs were also used in a permafrost environment. Aklavik is situated on a flat, silty clay terrace in the Mackenzie Delta. Some early twentieth -century buildings had conventional concrete basement walls that cracked and were displaced when ground ice beneath the buildings melted, causing differential settlement. Over time, the active layer deepened below buildings, trails, roads, and other structures, resulting in ground subsidence.

Because the presence of permafrost was not recognized, construction in permafrost terrain often used inappropriate practices, the consequences of which were later accepted or dealt with. Current engineering practices are more sophisticated and aim to minimize disturbance to the terrain as well as impacts on structures. Modern northern infrastructure is designed to preserve thaw-sensitive permafrost and therefore limit thaw settlement, located to avoid thaw-sensitive and frost-susceptible soils, or designed to withstand the anticipated thaw settlement or frost heave.

INFRASTRUCTURE DESIGN

Site investigations

Proper investigations of sites and routes in permafrost areas are central to obtaining the knowledge required for safe and sound infrastructure design, while ensuring minimal associated environmental impacts and avoiding unnecessary costs for construction, operation, and maintenance. The size and complexity of the project will determine the type and extent of site investigation required. For small projects, such as the construction of a single building, information on the ground thermal regime and the analysis of core samples from a few boreholes may be all that is required. For larger projects that traverse large distances and varied terrain, such as pipelines and highways, more comprehensive investigations may be required. These can include mapping of surficial materials and terrain features through field surveys and airphoto interpretation, sampling of soils from numerous boreholes, trenches, or surface exposures for analyses of physical, geotechnical, and thermal properties, monitoring ground temperatures to characterize the thermal regime, and conducting geophysical surveys (such as acoustic, electromagnetic, and ground-probing radar (Fig. 14) to map permafrost boundaries, delineate ice-rich zones or massive ground ice.

Infrastructure design commonly includes geothermal modelling to assess the terrain's thermal response to construction and operation of structures, and to determine if mitigative measures are required. Geothermal models, however, require a suite of data, including information on local climatic conditions, soil physical and thermal properties, and initial ground thermal conditions. For novel, complex, or critical structures, follow-up geothermal and geotechnical monitoring programs are generally established to evaluate performance, compare actual and predicted responses, and assess whether remedial work is required.

Transportation – roads, railways, and airfields

Roads, railways, and airfields in the permafrost region often cross widely varied types of terrain and may thus be subject to differential heave or differential thaw settlement. Problems can arise early in a project as the removal of vegetation and organic material along the route disturbs the ground thermal regime, causing an increase in summer ground temperatures and eventual thawing and settlement of ice-rich soil. The construction of a railway embankment or the placement of roadbed fill causes further changes in heat exchange at the ground surface. For example, the new, relatively dark surface, particularly if it is paved, may absorb more radiation during the summer than the natural surface. In addition, the cooling effect of evapotranspiration is greatly reduced when the vegetation cover is removed. As a result, an embankment or roadbed is usually warmer than the surrounding ground.

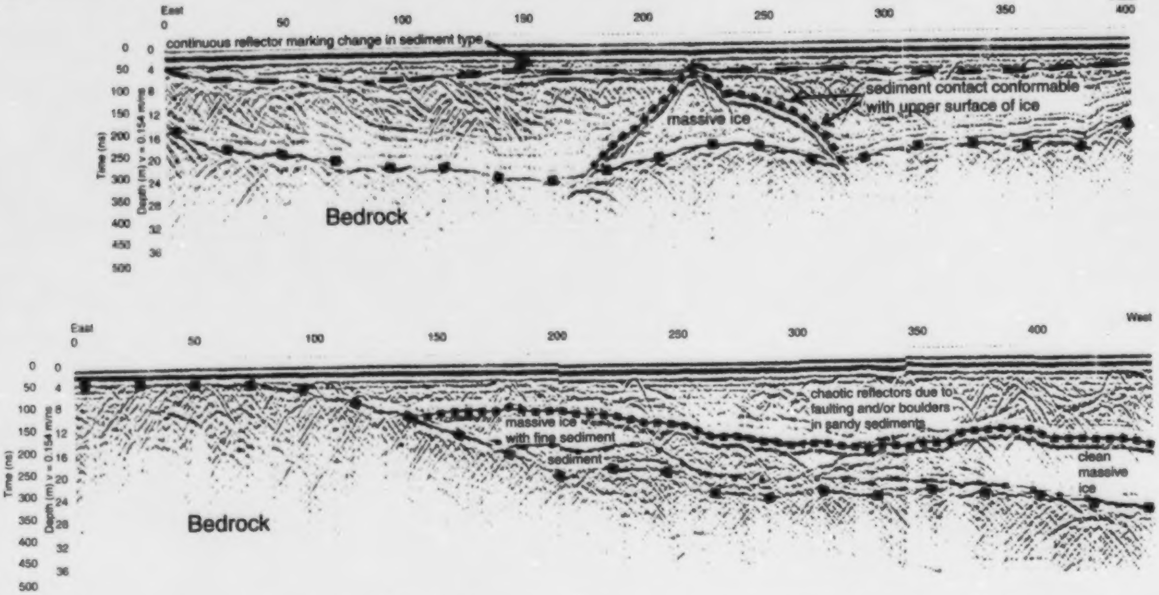


Figure 14. Ground ice in glaciofluvial deposits as interpreted from ground-probing-radar (GPR) profiles (after Wolfe et al., 1987).

Differential permafrost degradation and thaw settlement beneath a roadway can produce an uneven surface, as happened along initial sections of the Mackenzie Highway, in the western Northwest Territories. In winter, snow is usually cleared from roads or compacted by traffic, reducing its insulating effect. This results in deeper frost penetration and, where frost-susceptible soils are present, in increased frost heave. Constant road maintenance may be required, including regular grading and/or resurfacing, addition of fill, and raising and reconstruction of subsided curbs and sidewalks.

Similar problems can be encountered in railway construction. No special construction techniques were used in building the Hudson Bay Railway between The Pas and Churchill, Manitoba, in the early twentieth century. Between 1917 and 1926, practically no maintenance was undertaken on the line and severe subsidence from thawing occurred in many places (Fig. 15). This, combined with severe frost heave, twisted rails and even displaced some sections of track completely off the rail bed.

Route selection is very important with soil texture and drainage being the most important considerations. Fine-grained soils should be avoided wherever possible. In the discontinuous zone, where permafrost is patchy, it may be possible to avoid permafrost. Where permafrost is thin, the vegetation cover may be removed to promote thawing of permafrost. Frost-susceptible material can then be removed and replaced with nonfrost-susceptible granular material such as gravel.

Where permafrost is thicker and more extensive, its preservation is the prime consideration and insulation is used to retain the natural thermal regime as much as possible. This has been done, for example, by laying down a thick gravel bed under a roadway. Alternatively, a layer of thermal insulation,

Figure 15. Differential ground settlement (summer 1985) beneath a section of the railway used to provide access to dam sites along the Nelson River in northern Manitoba. This settlement was caused by melting of ice-rich permafrost that occurred in response to disturbance of the ground surface during railway construction. Photograph courtesy of E. Nielsen, Manitoba Geological Survey.

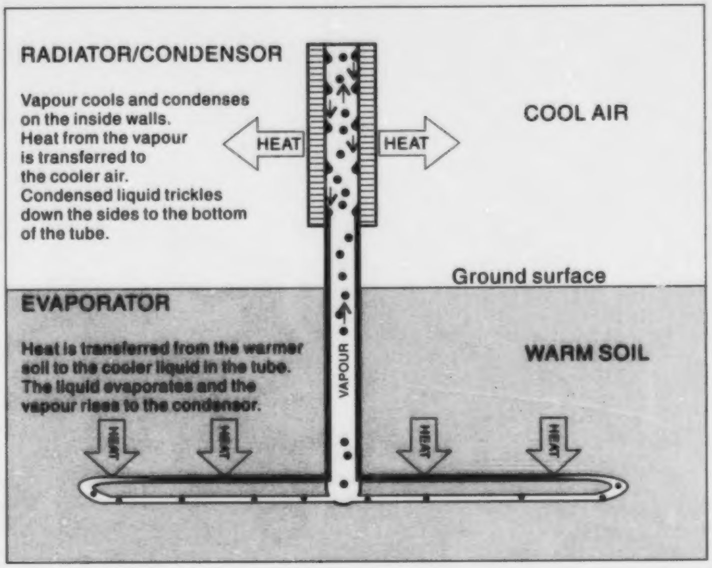

RADIATOR/CONDENSOR
Vapour cools and condenses on the inside walls. Heat from the vapour is transferred to the cooler air. Condensed liquid trickles down the sides to the bottom of the tube.
HEAT
HEAT
COOL AIR
Ground surface
EVAPORATOR
Heat is transferred from the warmer soil to the cooler liquid in the tube. The liquid evaporates and the vapour rises to the condensor.
WARM SOIL
VAPOUR
HEAT
HEAT
HEAT
HEAT

Figure 16.

A two-phase thermosyphon consisting of a sealed tube filled with a two-phase working fluid, such as carbon dioxide or anhydrous ammonia. It functions only in winter when the radiators are cold enough to cause condensation of the gas inside the tube. Continuous evaporation and condensation of the working fluid produces a very high rate of heat transfer from the ground to the air. The configuration shown here would be used to cool soil beneath a road. Thermosyphon tubes may also extend vertically below the ground surface as is the case for frozen-core dams (modified from *Wolfe, 1998).*

such as polystyrene, can be laid on the roadbed to reduce heat transfer to the permafrost. This method can be used when the required depth of fill to prevent thaw is large and aggregate is scarce. Polystyrene insulation was placed under a section of the Mackenzie Highway near Inuvik in 1972. By using at least 90 mm of insulation, thaw settlement was reduced to negligible levels and about half the thickness of fill was required compared to sections without insulation. Because some settlement is usually expected in the first few years following road construction, final surfacing is normally delayed until the new permafrost thermal regime has stabilized.

Similar techniques may be used for airfields, but normally much less differential settlement and cracking can be tolerated for airstrips than for roads. A smooth, finished surface is required for taxiing, takeoff, and landing. The airstrip in Inuvik was designed to prevent thawing of permafrost. In this case, an insulating layer of rock fill about 3 m thick was placed on undisturbed, ice-rich permafrost.

Thermosyphons (also called 'heat pipes', 'cryo-anchors', 'thermo tubes', or 'thermo piles') may be used to prevent permafrost from thawing beneath all-season roads. Thermosyphons are heat exchangers that, in winter, extract heat from the ground and transfer it to the air above by using a two-phase convective process (Fig. 16). The operation of thermosyphons in winter cools the permafrost and lowers its temperature sufficiently to prevent thawing in summer. In other words, the mean annual ground temperature is lowered during thermosyphon operation. Thermosyphons have been used in Yellowknife (Fig. 17) and elsewhere to lower the ground temperature beneath roads and parking lots. Thermosyphons have also been used to stabilize embankments along the aforementioned Hudson Bay Railway. They are self powered and require no electricity or other source of energy to operate, which gives them an advantage over other types of mechanical refrigeration systems.

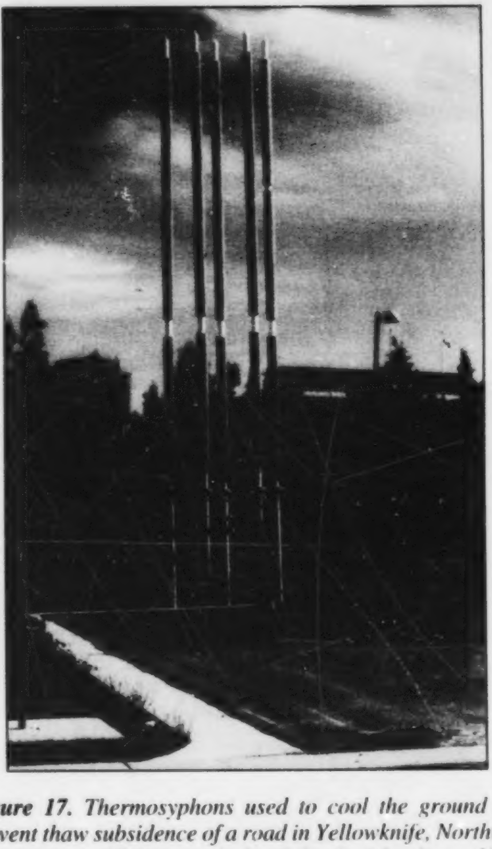


Figure 17. *Thermosyphons used to cool the ground and prevent thaw subsidence of a road in Yellowknife, Northwest Territories. Photograph by S.L. Smith, June 1998. GSC 2000-171*

Icings on road surfaces can also be a major problem in some permafrost areas. Icings are a seasonal buildup of ice up to 2 or 3 m thick that may block a roadway. Road icing generally arises in the fall and winter when seasonal freezing occurs at a faster rate beneath the road than the surrounding undisturbed area. The impermeable seasonal frost below the roadway prevents groundwater in the active layer on the uphill side of a road from moving downslope. This forces groundwater to the surface where it freezes, filling ditches and spreading over the road. The prevention of winter icing requires good drainage through the use of surface trenches and subsurface drains.

Winter roads still service many areas in the North. These seasonal roads rely on ground and water bodies freezing in winter to provide a stable, smooth surface. Winter roads are generally in service for only a few months of the year, as they require a certain amount of frost penetration for operation and maintenance. Examples of winter roads include the Mackenzie Highway from Wrigley to Fort Good Hope and from Inuvik to Tuktoyaktuk, and the road between Yellowknife and the Lupin gold mine at Contwoyto Lake, which also services the Ekati diamond mine at Lac de Gras.

Buildings and foundations

Frozen ground can provide excellent foundation support because ice forms a strong bond between the soil particles. However, problems can arise due to creep below building foundations in ice-rich soils and to the increase in soil temperature and thawing of ground ice. Differential thaw settlement of buildings may occur where thaw-unstable materials underlie only part of the building or where spatial variations occur in the heat produced by the building.

Different approaches can be taken to avoid problems of thaw settlement of buildings (Fig. 18). Where permafrost is shallow and patchy, such as in the southern margins of the discontinuous zone, the frozen soil may be excavated and replaced with non frost-susceptible material or intentionally thawed and allowed to settle prior to construction. This was done in Thompson, Manitoba, where surface vegetation was removed from housing lots that were then left vacant for several years until the permafrost disappeared. By following this procedure, damage to buildings was avoided.

In areas where permafrost is thicker and more extensive or continuous, intentional thawing of the permafrost and removal of thaw-sensitive material are not always feasible. If the foundation material consists of sound bedrock or dense, thaw-stable, granular soils, such as clean, well drained sand and gravel, conventional design techniques used in nonpermafrost zones may be used. Thus, site selection is important. For example, when the federal government initiated construction of the town of Inuvik in 1954, site selection was based primarily on the frost stability of the underlying material. The Inuvik townsite is on high ground and underlain by coarse-grained soil with good drainage and minimal ground-ice content. A large body of fluvioglacial gravel near the townsite provided suitable aggregate for foundation construction. Inuvik is an example of how careful planning can reduce and avoid problems otherwise encountered in large-scale construction in permafrost areas.

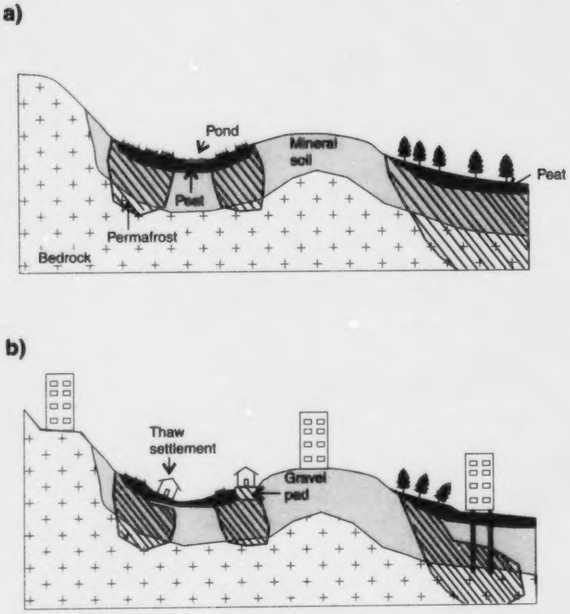


Figure 18. a*) Natural permafrost conditions for a region within the discontinuous permafrost zone.* ***b****) The impact of building construction on this terrain and techniques used to minimize the effects of thaw subsidence. Construction methods used in nonpermafrost areas have also been used successfully in areas where permafrost is absent. If no special precautions are taken in areas of ice-rich permafrost, thaw settlement of the ground and building can occur. Insulation provided by a gravel pad, for example, helps to preserve permafrost beneath the building and prevent thaw settlement. Buildings overlying permafrost can also be constructed on piles anchored into bedrock to prevent settlement of the building. However, thawing of the permafrost and settlement of the ground beneath and around the building may still occur. Note: diagram not to scale.*

Where construction must take place on thaw-unstable permafrost, techniques can be used to preserve the permafrost while allowing freezing and thawing of the active layer. Permafrost under a building may be preserved by using insulation (similar to that described for roads), ventilation, or a combination of both. Natural ventilation can be provided by constructing a building above the ground on piles or other supports (Fig. 19). In winter, the flow of cold air under the building dissipates the heat produced by the structure and maintains cold ground temperatures. Shade provided by the building in summer and the lack of snow in winter combine to cool the ground over time and prevent or reduce thawing of permafrost. Duct-ventilated gravel pads are used for buildings such as aircraft hangers, maintenance garages, or storage

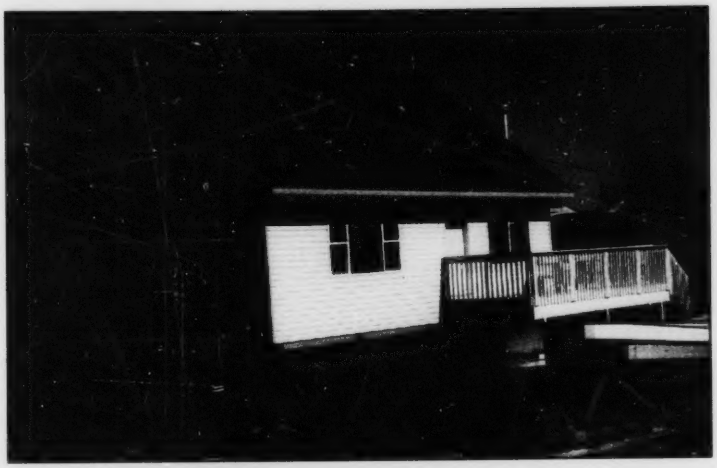


Figure 19.

House constructed on screw jacks (enlargement shown in inset) in Norman Wells, Northwest Territories. The screw jacks are used to facilitate levelling of the house following periodic differential thaw settlement or heave. Elevation of the building above the ground surface provides ventilation and allows dissipation of the heat produced by the house. Photograph by M.M. Burgess, July 1998. GSC 2000-172

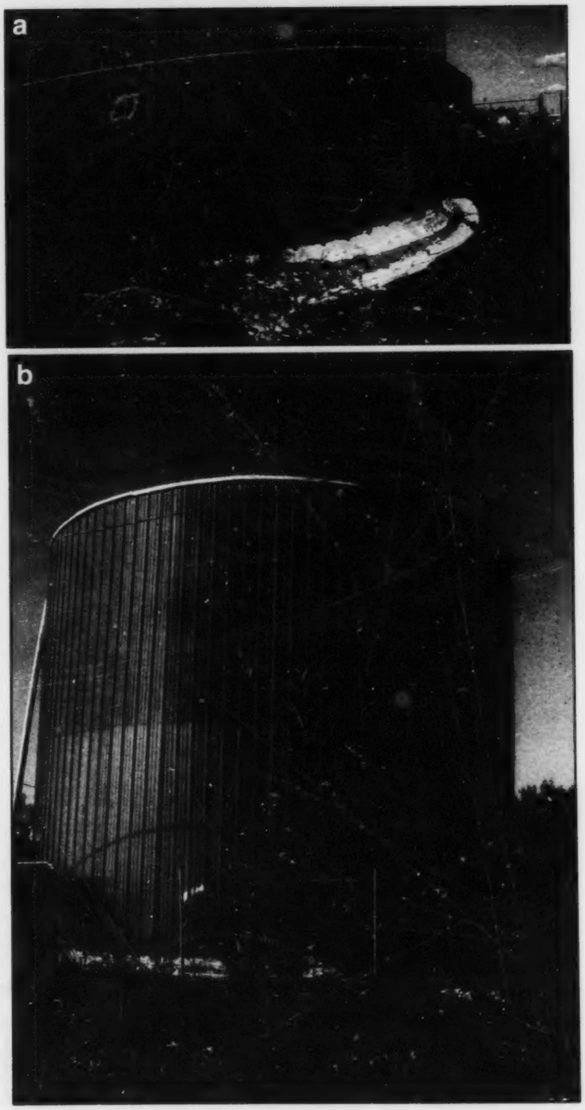


***Figure 20. a**) Cryo-anchors (an early type of thermosyphon) installed in the late 1970s to prevent thaw subsidence beneath **b**) a water storage tank in Norman Wells. Photographs by M.M. Burgess, June 1998. **a**) GSC 2000-173; **b**) GSC 2000-174*

tanks that must be placed on grade for access and/or to distribute large loads. Natural convection or fans are used to circulate cold winter air through ducts in a gravel pad to dissipate the heat produced by the structure. In summer, the ducts are blocked to reduce warming and retard thawing.

Thermosyphons have also been used to cool the ground beneath buildings and other structures and to help stabilize foundations (Fig. 20). Thermosyphon evaporator tubes were used in winter to freeze the gravel pad beneath a school in Ross River, Yukon Territory. The annual thawing and freezing is confined to the fill, thus preserving the underlying permafrost. No evidence has been found of significant heave or settlement of the building or cracking of the concrete floor slab.

Nevertheless, some thaw and ground settlement may occur beneath buildings due to ground disturbance. With small, simple structures, some settlement can be tolerated. Periodic jacking and levelling of the building may be required in more extreme cases of settlement. Houses, for example, can be built on a system of jacks to facilitate adjustments following differential movements (Fig. 19). Larger, more complex structures are often placed on piles anchored into bedrock or deep in permafrost, thus preventing the building from settling whether or not settlement of the underlying soil occurs (Fig. 21).

Frost action in the active layer or within the seasonally frozen ground in the discontinuous permafrost zone can subject foundations and structures to large uplift forces. Frost heave may have to be considered in the design of foundations for buildings or other structures (e.g. power transmission towers), whether on piles or on gravel pads. Upward displacement or frost jacking of structures may result from vertical forces acting at the base of the foundation or pile or from adfreeze (the process by which two objects are bonded

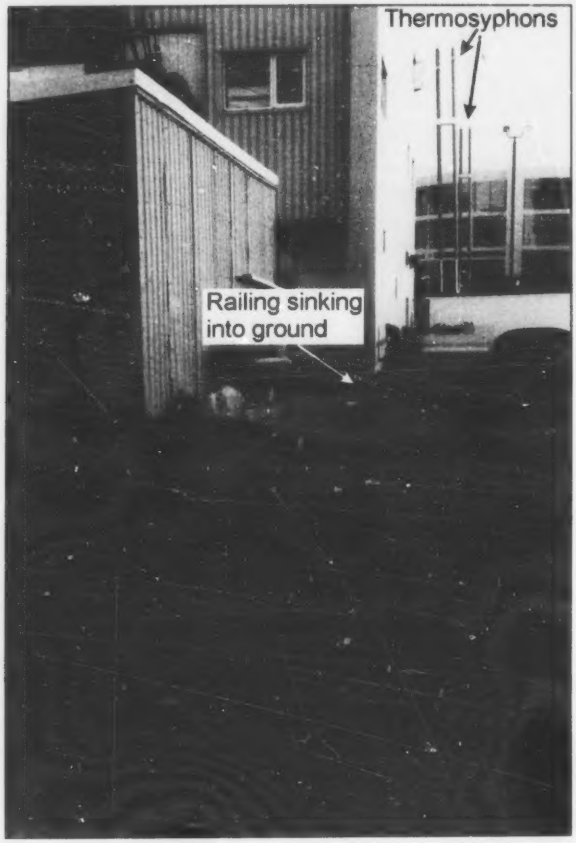


Figure 21. Thaw subsidence beneath and around a heated building in Yellowknife, Northwest Territories, has resulted in sinking of the steel guardrail beside the building. The building is built on piles anchored in bedrock and thus has remained stable. Fill has been used to raise the ground surface back to grade. Thermosyphons used to cool the ground beneath an adjacent road can be seen in the background. Photograph by M.M. Burgess, June 1998. GSC 2000-175

together by ice formed between them) forces produced by the growth of ice lenses in the soil adjacent to the footings or piles. These forces can act differentially, causing severe distortion of structures. Over several years, this displacement will be permanent if the heaved structure cannot return to its original position at the end of each thaw season (Fig. 9). Thawing of ice lenses beneath foundations in the spring can cause a decrease in bearing capacity and stability problems, particularly if it occurs faster than moisture can drain away.

Controlling frost heave involves the elimination of one or more of the conditions required for ice-lens formation, i.e. frost-susceptible material, a supply of water, and ground temperatures below 0°C. One solution is to replace frost-susceptible material in the active layer with coarser grained material during foundation construction. Providing good drainage will reduce the supply of water for ice-lens growth. Seasonal frost penetration may also be reduced by placing insulation at the ground surface. Heave at the base of foundations or piles can be reduced by placing them below the depth of seasonal frost or below the permafrost base. Heave due to adfreeze forces, however, may still be a problem. Anchoring piles (for example in bedrock) or embedding them deeply in permafrost will reduce heave due to adfreeze forces. Piles can also have sleeves placed around them to isolate them from the heaving ground and eliminate or reduce the adfreeze forces.

Utilities

In southern communities, water and sewage pipes are installed below the depth of seasonal frost penetration to prevent the pipes from freezing. Although this can be done in nonpermafrost areas within the discontinuous permafrost zone, it is impossible where permafrost is present. In permafrost areas, pipes placed within the active layer can be used in summer, but will freeze in winter. Several solutions are available to address this problem. Often the simplest and most cost-effective solution is to provide piped water and sewage disposal to the community in summer, and to supply drinking water and remove sewage by truck from individual houses and buildings in winter. This method is used in the older section of Yellowknife and at Sachs Harbour, Banks Island. In new subdivisions in Norman Wells, Northwest Territories, water and sewage are trucked year round.

The freezing of buried water pipes can be prevented by maintaining a continuous flow of water or by heating. Sewer pipes can also be prevented from freezing by being located close to heated waterlines. Insulation around pipes can also prevent freezing. Of course, ground ice can thaw, leading to thaw settlement, where heated pipes are placed in thaw-unstable material. This can cause pipes to break, especially at boundaries between permafrost and nonpermafrost areas or between thaw-stable and thaw-unstable soil where differential settlement may occur (Fig. 22). Insulating pipes can limit thaw, and trenching and backfilling above and below pipes with sand or gravel can minimize settlement.

Utilidors can be used in major communities where permafrost is widespread and thick and the active layer is shallow. Utilidors are heavily insulated enclosures, located above or below the ground surface, that contain utility pipes. They may be heated and may also carry steam or hot water heating pipes, fuel distribution lines, and electrical cables, as well as water and sewage mains. They are expensive and are not normally used except where alternative methods cannot be used. To minimize disturbance to the ground thermal regime, utilidors usually are placed on piles or on gravel pads. They may also be buried in shallow trenches either in the active layer or in thaw-stable frozen material. They must, however, be designed to prevent thermal disturbance to the underlying permafrost and also be constructed to prevent infiltration of groundwater or surface water. Both shallow underground utilidors and elevated utilidors on piles driven into bedrock are currently used for water and sewage in parts of Norman Wells (Fig. 23). Utilidors are also used within the continuous permafrost zone, in Inuvik for example, where they also provide district heating, which helps prevent water and sewage lines from freezing in winter. Should the town eliminate district heating, an alternative means of preventing or reducing the freezing of water and sewage lines will be required.

Locating a sufficient supply of good quality drinking water is commonly a problem for communities in the permafrost region. Where available, surface water bodies must not only be deep enough to prevent freezing to the bottom in winter, but of a suitable quality for drinking water. Groundwater is another option, but its supply is commonly limited, especially in the continuous permafrost zone. Groundwater movement is restricted to taliks because permafrost acts as an impermeable layer. Groundwater above permafrost is usually present only in small volumes and is commonly unsuitable for human consumption. Groundwater may occur within permafrost, but it is commonly mineralized. Groundwater beneath the permafrost base is the most promising source of water, but drilling through permafrost over 100 m thick can be difficult and expensive. In addition, wells must operate continuously to prevent freezing of the pipes or the wellbore must be heated, which can induce ground-stability problems in ice-rich soil. Wrigley, in the Mackenzie River valley, is an example of a community in the Northwest Territories that uses subpermafrost groundwater. The water supply there has been used successfully for about 20 years, despite permafrost being 40 to 50 m thick.

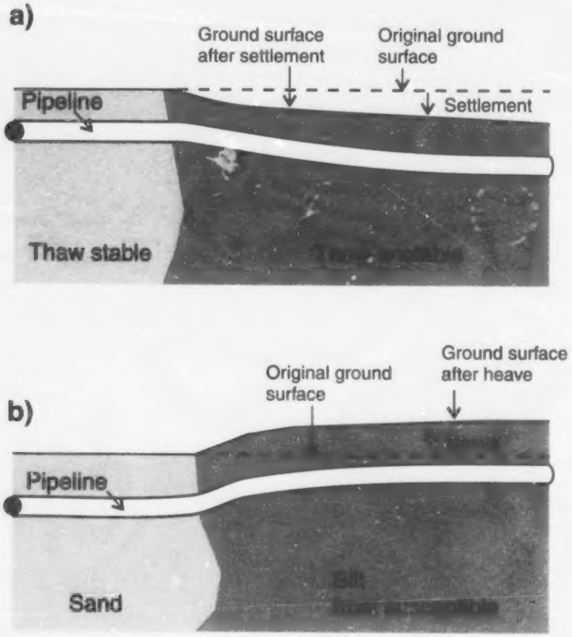


Figure 22. a*) Differential settlement of a warm pipeline at a transition between thaw-stable and thaw-unstable material.* ***b****) Differential heave of a chilled pipeline at a transition between frost-susceptible silt and nonfrost-susceptible sand.*

Pipelines

One of the major challenges facing the construction and operation of buried pipelines in the North is the many terrain types, materials, and permafrost conditions that must be traversed. If the pipeline operates at temperatures above freezing, it can cause melting of permafrost and thaw subsidence. In addition, the disturbance to the ground thermal regime caused by removal of vegetation and organic material along the pipeline right-of-way will result in permafrost degradation. Pipe stability may be affected where differential thaw settlement occurs at terrain transitions. Where the pipeline traverses slopes, thawing of permafrost may lead to slope instability. The pipeline can be chilled to prevent permafrost thaw, but in the discontinuous permafrost zone, subsequent freezing and frost heave can occur in nonpermafrost soils. Differential heave or settlement of the pipe and ground surface can occur at boundaries between permafrost and unfrozen ground and between frost-susceptible (thaw-unstable) and nonfrost- susceptible (thaw-stable) materials (Fig. 22).

The Norman Wells pipeline, owned and operated by Enbridge Pipelines Inc. (formerly Interprovincial Pipe Line), is the first completely buried oil pipeline in the Canadian permafrost zone. Extending from Norman Wells, Northwest

Figure 23.

Utilidor in Norman Wells, Northwest Territories. This contains water-supply and sewage pipes for part of the community and is supported by piles. The building in the background on the left is part of a recreational complex and is also supported by piles. Photograph by M.M. Burgess, July 1998. GSC 2000-176

Territories, to Zama, northwestern Alberta, the 869 km pipeline route follows a north-south transect through the discontinuous permafrost zone, from extensive discontinuous permafrost in the north to sporadic permafrost in the south. The pipeline has been in operation successfully since April 1985. Preconstruction geotechnical investigations and design studies included a continuous geophysical survey along the right-of-way to map the extent of frozen ground and locate frozen-unfrozen interfaces; mapping of surficial geology at a scale of 1:20 000; thaw-depth and settlement surveys and analyses; drilling of several hundred geotechnical boreholes; and geothermal modelling. Geotechnical issues considered in the pipeline design included the thermal mode of operation, thaw settlement, frost heave, slope stability, and drainage and erosion control. Fine-grained lacustrine soils as well as frozen organic soils along the pipeline route are commonly ice rich and hence thaw sensitive. As a result, a number of unique design features and mitigative measures were developed to minimize terrain disturbance and ensure pipeline integrity. The key features are described below.

The pipeline was designed as a small-diameter (328 mm), ambient-temperature pipeline operating at temperatures within a few degrees of the surrounding soils to prevent the transfer of significant heat into the ground. To achieve this, the oil is chilled at Norman Wells and enters the pipeline at a mean annual temperature of about -1.0°C. Right-of-way clearing generally took place one year prior to construction, to allow for some of the related thaw settlement to occur prior to trenching and pipeline burial. Whereever possible, the right-of-way followed existing clearings or cutlines to benefit from the associated thaw settlement that had already occurred. To minimize surface disturbance and reduce costs, construction was undertaken over two winters using temporary snow roads (Fig. 24); no permanent workpad or road was associated with the project.

Removal of vegetation and soil disturbance along the right-of-way were predicted to cause a net warming of the ground surface. The associated thaw settlement in ice-rich soils presented a concern for the integrity of the pipe, particularly where differential movements were likely to occur across sharp soil textural and/or thermal interfaces. The pipeline was designed to withstand the likely total and differential thaw settlements expected to occur over its 25 to 30 year design life.

The section of the pipeline route (Norman Wells to Fort Simpson) that follows the east bank of the Mackenzie River crosses numerous tributary rivers and creeks. Particular attention was paid to the design of permafrost slopes at these crossings to ensure slope stability. Detailed, site-specific designs were developed, with 56 permafrost slopes receiving an insulating cover of wood chips up to 1.5 m thick (Fig. 25). Recognizing the uniqueness of many aspects of the project, both the federal government and Enbridge Pipelines Inc. established geotechnical and thermal monitoring programs along the pipeline. For further information on these programs and on the pipeline performance, *see* 'Suggested reading'.

Dams and dykes

Problems of seepage, stability, and slope protection must be considered when constructimng dams and dykes in both permafrost and nonpermafrost regions. In the permafrost region, there is the additional thawing effect of the impounded water on frozen material underlying the reservoir and the dam or dyke structure. The design must also consider the thermal regime of the foundation material as well as the embankment itself in the case of earthfill structures. Differential settlement of a dam or dyke could produce unacceptable or uncontrolled seepages that threaten the stability of the structure, possibly causing catastrophic failure. Depending on the purpose of the structure, failure could result in serious impacts on the surrounding environment such as flooding and erosion or the release of hazardous material from tailings ponds or sewage lagoons.

Frozen earthfill dams are often used in the permafrost region to take advantage of the high strength and impermeable nature of frozen ground. These can be made of any kind of soil, and the foundation and core of the dam are frozen during or immediately after construction. The fill must be saturated with ice to produce a well bonded mass and this condition must be maintained throughout the life of the structure.

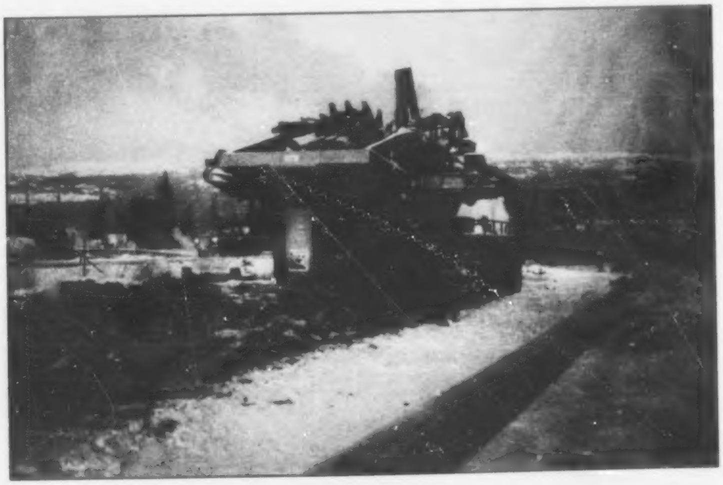


Figure 24.

Special arctic ditchers in operation during construction of the Norman Wells pipeline in 1984. The ditchers were designed to cut a 1.2 m deep trench through the frozen active layer and into the permafrost soils. The pipe is laid out on the travel side of the right-of-way, i.e. the side used as a temporary winter road during construction. The vehicles in the background are on an all weather road outside Norman Wells. Photograph taken in January 1984. GSC 2000-033A

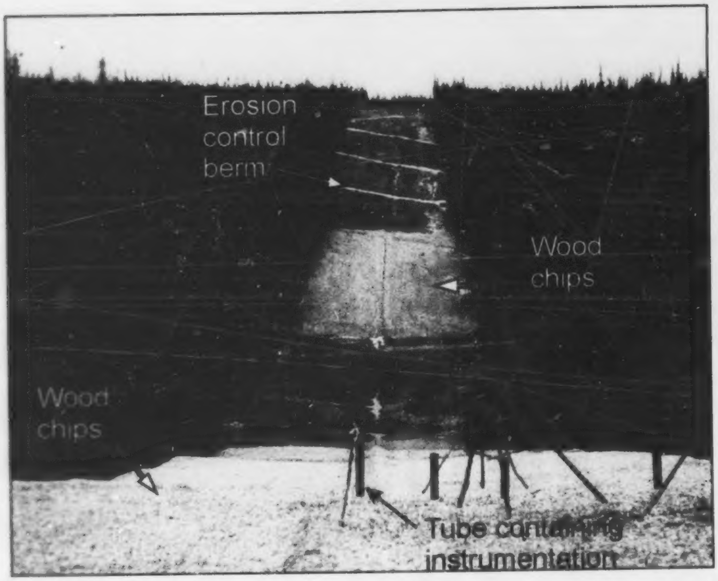


Figure 25.

Section of the Norman Wells pipeline at Canyon Creek in 1984, the first summer after construction. The slopes of the creek valley are insulated with a 1 m layer of wood chips to maintain the stability of the underlying permafrost. The lines cutting diagonally across the slope in the background are erosion-control sand berms. The black tubes in the foreground contain instrumentation used for monitoring subsurface temperature. Photograph by M.M. Burgess, June 1984. GSC 2000-177

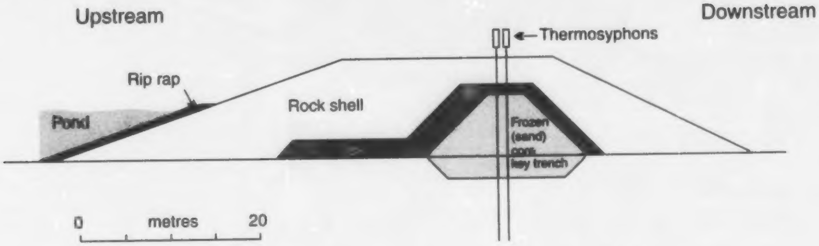


Figure 26. *Design of frozen-core dam for the Ekati diamond mine, Northwest Territories (from BHP Diamonds Inc, 1995). Thermosyphon tubes are used to chill the dam and the underlying ground in winter to maintain the impermeable frozen core.*

The fill is built up in layers, each of which is allowed to freeze naturally. This method was used to construct a tailings dam at the Polaris mine on Little Cornwallis Island, Nunavut, in the continuous permafrost zone. The dam was constructed in the winter of 1990–1991 by spreading and compacting 30 to 40 cm of fill each day; this fill then froze overnight. Styrofoam insulation was placed on the surface of the dam to reduce thaw during the summer. At the Lupin gold mine, Nunavut, the tailings dams were also built on frozen ground. In this case, the fill froze permanently the first year, providing a watertight structure. Care is taken to maintain water levels in the pond below the depth of the active layer of the dam. Similar techniques were used in the design of dykes for containing acid-generating waste rock at the Raglan nickel mine in northern Quebec.

For larger earthfill dams, the core is normally artificially frozen by mechanical refrigeration, the circulation of cold air through vertical ducts, or the use of thermosyphons. Thermosyphons have been used at the Ekati diamond mine site, in the Northwest Territories, in frozen-cored tailings and water-diversion dams (Fig. 26, 27). The tailings pond dam is partly situated on a talik and refrigeration is required to prevent seepage and instability.

In the discontinuous zone, thawing of soils below dams and dykes is a major concern. In the case of the Kelsey generating station on the Nelson River in northern Manitoba, the earthfill dykes were placed directly on permafrost and excess water was removed from the thawing soil by sand drains installed beneath the dykes. The soil below the dykes was then allowed to consolidate naturally, and new fill was added to bring the crests to the required elevation. The dykes have performed well and no major instability problems have occurred. However, the impoundment of water and the flooding of permafrost terrain have led to thawing of permafrost along the reservoir shoreline. Erosion and solifluction have resulted in slumping and increased turbidity in the lake.

PERMAFROST AND CLIMATE CHANGE

In recent years, climate change has been recognized as an additional challenge to development in permafrost regions. Permafrost is a thermal condition, the very existence of which is dependent on climate. Climate, however, is not constant. It has undergone changes detectable at time scales from

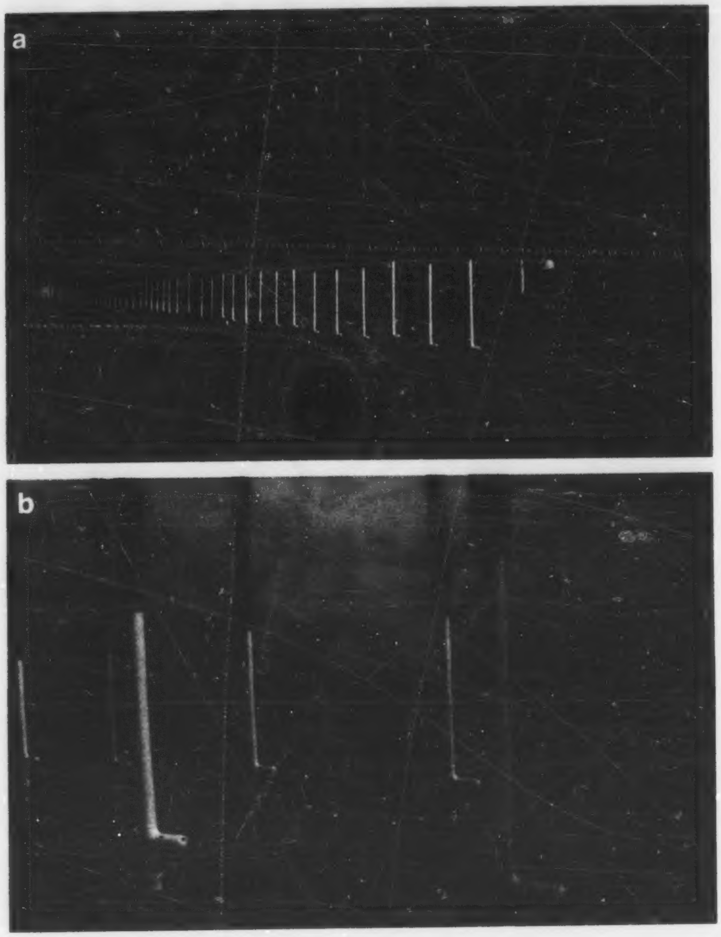


Figure 27.

***a)** Thermosyphon tubes extending from the surface of the frozen-core dam at the Ekati diamond mine site, Northwest Territories. Thermosyphon tubes are used to chill the dam and the underlying ground in winter to ensure that the core is kept frozen. Photograph by M.M. Burgess GSC 2000-178* ***b)** The black tube contains instrumentation for monitoring the internal temperature of the dam. Photograph by M.M. Burgess, June 1998 GSC 2000-179*

decades or centuries to millennia. As climate cools, permafrost increases in areal extent and thickens; a warmer climate causes thickening of the active layer as well as thinning and in some cases disappearance of permafrost. The surface buffer layer and the natural damping effect of subsurface materials cause a lag in the permafrost response to changes in surface conditions or climate.

Not all permafrost is in equilibrium with today's climate. As discussed under 'Permafrost distribution and ground thermal regime', thick permafrost under the Beaufort Shelf was formed during periods of lower sea level associated with the last glaciation. Frozen peatlands occur today in the southern fringe of the discontinuous zone at relatively warm ground temperatures (above -0.5°C). Permafrost in these peatlands likely formed when slightly colder climatic conditions prevailed in the northern hemisphere during a period known as the 'Little Ice Age', which ocurred between 100 and 600 years ago. This permafrost has been preserved because of the insulating properties of the thick peat cover. Between 6000 and 9000 years ago, climate was warmer than today and the southern limit of permafrost was north of its present position, and much of the present discontinuous permafrost zone may have been free of permafrost.

Permafrost is also dynamic in today's climate. Interannual variations in rainfall and snowfall or the occurrence of drought or fires causes changes in the permafrost environment. A dramatic illustration of this is the marked increase in active-layer detachment slides along the banks of the Mackenzie River following forest fires near Tulita in the warm spring of 1995 (Fig. 12). The local distribution and extent of permafrost also varies in response to changes in the surface thermal regime caused, for example, by the lateral migration of river channels, shoreline erosion, lake drainage, and coastal emergence.

Extreme climatic events, such as higher than normal air temperatures associated with El Niño events like those of 1998, may also affect the permafrost environment. In the western Northwest Territories, the average air temperature for 1998 was the highest since 1948. Values were almost 4°C above normal with the greatest warming occurring in spring and winter. Events such as this may well lead to increased active-layer thickness, thaw settlement, and slope instability.

The international Intergovernmental Panel on Climate Change (1966) declared that "the balance of evidence suggests a discernible human influence on global climate."

Atmospheric carbon dioxide (CO_2) concentration is expected to have doubled by 2050–2100, and CO_2 levels will likely increase beyond that, especially if greenhouse-gas emission-reduction targets are not met. General circulation models predict that for a doubling of the atmospheric content of CO_2, average global air temperature could rise by 1 to 3.5°C. Warming is expected to be greater in the North where mean annual air temperatures could increase up to 6 or 7°C due to feedback effects from the polar ice pack. Increases in air temperature will lead to an increase in permafrost temperature, although changes in temperature below the ground surface would generally be less than those above. The buffer layer, which includes the vegetative, organic, and snow covers, plays an important role, and changes to the buffer layer associated with climate warming will greatly influence the magnitude and rate of permafrost response. Local ground-ice and ground thermal conditions are also important, with warm permafrost (warmer than -2°C) being the most vulnerable to climate change. The results of geothermal modelling from the Mackenzie River valley indicate that permafrost response to warming is slower at sites with ice-rich soils. The modelling also predicts that complete degradation of permafrost may be possible within the next 50 years in the southern discontinuous permafrost zone of the Mackenzie River valley near Fort Simpson, where permafrost is only a few metres thick. Warming of ice-rich soils will proceed at a slower rate than that of ice-poor soils, although the physical impact of associated thaw will be greater. In the northern, continuous permafrost regions, the active layer will increase in thickness, permafrost temperatures will increase, and permafrost thickness will slowly decrease.

Warming and thawing of permafrost due to climate change may result in changes to the engineering behaviour of soils and instabilities in the landscape, such as an increase in thermokarst or a greater incidence of active-layer detachment slides, particularly if the number of forest fires increases. These in turn will affect northern ecosystems, communities, infrastructure, and lifestyles. Increased thaw depths and thaw settlement may necessitate increased maintenance of infrastructure elements, and concerns for the integrity of structures may lead to calls for remedial and/or adaptive work.

In 1981, the large uncertainty surrounding global warming was such that climate change was not considered in the design of the Norman Wells pipeline. Despite continued uncertainty about the magnitude and rate of climate warming, especially at a regional scale, climate warming was recognized as being a concern over the life of the new Ekati diamond mine in the Northwest Territories, which opened in late 1998. The potential impact of warming was considered in the design of the mine's frozen-core tailing dams.

ACKNOWLEDGMENTS

Comments on the manuscript by Michael Smith of Carleton University and Larry Dyke and Greg Brooks of the Geological Survey of Canada are greatly appreciated. Janice Naufal of the Geological Survey of Canada provided digital versions of the photographs. Daniel T. Desrochers of Environmental Analysis Services assisted in the production of the maps.

REFERENCES

Atmospheric Environment Service
1982: Temperature 1951–1980; Canadian Climate Normals; Atmospheric Environment Service, Environment Canada, v. 2, 306 p.
Bostock, H.S.
1970: Physiographic regions of Canada; Geological Survey of Canada, Map 1254A, scale 1:5 000 000.
BHP Diamonds Inc.
1995: NWT Diamonds Project, Tailings management plan and preliminary design of retention structures; report submitted by EBA Engineering Consultants Ltd., 107 p.
Brown, R.J.E.
1978: Permafrost; *in* Hydrological Atlas of Canada, Plate 32; Fisheries and Environment Canada.
French, H.M.
1996: The Periglacial Environment (second edition); Longman Group Ltd., New York, 341 p.
Heginbottom, J.A. and Radburn, L.K. (comp.)
1992: Permafrost and ground ice conditions of Northwestern Canada; Geological Survey of Canada, Map 1691A (two sheets), scale 1:1 000 000.
Heginbottom, J.A., Dubreuil, M-A., and Harker, P.A. (comp.)
1995: Canada — Permafrost; National Atlas of Canada (fifth edition), Plate 2.1; Geomatics Canada, National Atlas Information Service, and Geological Survey of Canada, MCR 4177, scale 1:7 500 000.
Intergovernmental Panel on Climate Change
1996: The science of climate change; Contribution of Working Group I to the Second Assessment Report of the Intergovernmental Panel on Climate Change, 584 p.
International Permafrost Association
1998: Multi-language glossary of permafrost and related ground ice terms, version 2.0; (ed.) R.O. vanEverdingen, University of Calgary, Calgary, Alberta, 278 p.
Permafrost Subcommittee
1988: Glossary of permafrost and ground ice terms; National Research Council of Canada, Associate Committee on Geotechnical Research, Technical Memorandum No. 142, 156 p.
Wolfe, S.A. (ed.)
1998: Living with frozen ground. A field guide to permafrost in Yellowknife, Northwest Territories; Geological Survey of Canada, Miscellaneous Report 64, 71 p.
Wolfe, S.A., Burgess, M.M., Douma, M., Hyde, C., and Robinson, S.D.
1997: Geological and geotech;nical investigations of massive ice in glaciofluvial deposits, Slave Geological Province, N.W.T.; Geological Survey of Canada, Open File 3442, 49 p.

SUGGESTED READING

General

Brown, R.J.E.
1970: Permafrost in Canada: Its Influence on Northern Development; Toronto, University of Toronto Press, 234 p.
Burn, C.R. and Smith, M.W.
1993: Issues in Canadian permafrost research; Progress in Physical Geography, v. 17, p. 156–172.
Carter, L.D., Heginbottom, J.A., and Woo, M-k.
1987: Arctic Lowlands; *in* Geomorphic Systems of North America, (ed.) W.L. Graf; Geological Society of America, Decade of North American Geology, Centennial Special Volume 2, p. 583–628.
French, H.M. (ed.)
1986: Permafrost geomorphology; Canadian Geographer, v. 30, p. 358–366.
Heginbottom, J.A. (co-ord.)
1989: A survey of geomorphic processes in Canada; Chapter 9 *in* Quaternary Geology of Canada and Greenland, (ed.) R.J. Fulton; Geological Survey of Canada, Geology of Canada, no. 1, p. 573–644 (*also* Geological Society of America, The Geology of North America, v. K-1), p. 573–644.

International Permafrost Association
1998: Permafrost Bibliography 1978–1997; CAPS Version 1.0 CD-ROM, published by the National Snow and Ice Data Centre, Boulder, Colorado.

Johnston, G.H. (ed.)
1981: Permafrost — Engineering, Design and Construction; Toronto, John Wiley and Sons, Toronto, Ontario, 540 p.

Smith, D.W. and Low, N. (ed.)
1996: Cold regions utilities monograph (third edition); Technical Council on Cold Regions Engineering, American Society of Civil Engineers and Cold Regions Engineering Division, Canadian Society for Civil Engineering.

Washburn, A.L.
1979: Geocryology; Edward Arnold, London, England, 406 p.

Williams, P.J.
1979: Pipelines and Permafrost: Physical Geography and Development in the Circumpolar North; Longman, London, England, 98 p.

Williams, P.J. and Smith, M.W.
1989: The Frozen Earth: Fundamentals of Geocryology; Cambridge University Press, Cambridge, United Kingdom, 306 p.

Maps

Brown, J., Ferrians, O.J., Jr., Heginbottom, J.A., and Melnikov, E.S. (comp.)
1993: Circumarctic map of permafrost and ground ice conditions; United States Geological Survey, Circum-Pacific Map Series, Map CP-45, scale 1:10 000 000.

Fulton, R.J.
1995: Surficial materials of Canada; Geological Survey of Canada, Map 1880A, scale 1:5 000 000.

Heginbottom, J.A. and Dubreuil, M-A.
1993: A new permafrost and ground ice map for the National Atlas of Canada; *in* Proceedings, Sixth International Conference on Permafrost, v. 1, p. 255–260.

Heginbottom, J.A., Bown, J., Melnikov, E.S., and Ferrians, O.J., Jr.
1993: Circumarctic map of permafrost and ground ice conditions; *in* Proceedings, Sixth International Conference on Permafrost, v. 2, p. 1132–1136.

Tarnocai, C., Kettles, I.M., and Ballard, M.
1995: Peatlands of Canada; Geological Survey of Canada, Open File 3152, scale 1:6 000 000.

Engineering geology

Andersland, O.B. and Anderson, D.M. (ed.)
1978: Geotechnical Engineering for Cold Regions; McGraw-Hill Book Company, New York, 566 p.

Andersland, O.B. and Ladanyi, B.
1994: An Introduction to Frozen Ground Engineering; Chapman and Hall, New York, 352 p.

Cathro, D.C., Keen, A.J., and Haley, D.W.
1992: Design and construction of a permafrost core earth dam; *in* Mining in the Arctic, (ed.) S. Bandopadhyay and M.G. Nelson; *in* Proceedings, Second International Symposium on Mining in the Arctic, Fairbanks, Alaska, p. 153–162.

Cruden, D.A., Bornhold, B.D., Chagnon, J.-Y., Evans, S.G., Heginbottom, J.A., Locat, J., Moran, K., Piper, D.J.W., Powell, R., Prior, D., Quigley, R.M., and Thomson, S.
1989: Landslides: extent and economic significance in Canada; *in* Landslides: Extent and Economic Significance, (ed.) E.E. Brabb and B.L. Harrods; Proceedings, 28th International Geological Congress: Symposium on Landslides, Washington, D.C., July 17, 1989, p. 1–23.

Dufour, S. and Holubec, I.
1988: Performance of two earthfill dams at Lupin, N.W.T.; *in* Proceedings, Fifth International Permafrost Conference, v. 2, p. 1217–1222.

Giegerich, H.M.
1988: Arctic mining in permafrost; *in* Proceedings, Fifth International Permafrost Conference, v. 2, p. 1382–1387.

Goodrich, L.E.
1983: Thermal performance of a section of the Mackenzie Highway; *in* Proceedings, Fourth International Permafrost Conference, p. 353–362.

Graham, C.A.
1986: Impact of offshore permafrost on oil and gas production; *in* Proceedings, Workshop on Subsea Permafrost and Pipelines in Permafrost; National Research Council of Canada, Technical Memorandum no. 139, p. 35–47.

Hanna, A.J., Forsyth, R.J., and Garvin, D.
1990: Thaw settlement around a building on warm ice-rich permafrost; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana no. 54, p. 419–424.

Hayley, D.W.
1982: Application of heat pipes to design of shallow foundations on permafrost; *in* Proceedings, Fourth Canadian Permafrost Conference, p. 535–544.

Hayley, D.W., Roggensack, W.D, Jubien, W.E., and Johnson, P.J.
1982: Stabilization of sinkholes on the Hudson Bay Railway; *in* Proceedings, Fourth International Permafrost Conference, p. 468–473.

Heuer, C.E., Caldwell, J.B., and Zamsky, B.
1982: Design of buried seafloor pipelines for permafrost thaw settlement; *in* Proceedings, Fourth International Permafrost Conference, p. 486–491.

Johnston, G.H.
1982: Design and performance of the Inuvik, N.W.T., airstrip; *in* Proceedings, Fourth Canadian Permafrost Conference, p. 577–585.
1982: Performance of insulated roadway on permafrost, Inuvik, N.W.T.; *in* Proceedings, Fourth International Permafrost Conference, p. 548–553.

Klassen, H.P.
1965: Public utilities problems in the discontinuous permafrost areas; *in* Proceedings, Canadian Regional Permafrost Conference; National Research Council of Canada, Technical Memorandum no. 86, p. 106–117.

Lafleche, P.T., Judge, A.S., and Pilon, J.A.
1988: The use of ground penetrating radar in the design and monitoring of water retaining embankments in permafrost; *in* Proceedings, Fifth International Permafrost Conference, v. 2, p. 971–976.

Mitchell, D.E., Curtis, D.D., Laut, S.W., Pui, N.K., and Burgess, A.S.
1983: Well casing strains due to permafrost thaw subsidence in the Canadian Beaufort Sea; *in* Proceedings, Fourth International Permafrost Conference, p. 855–860.

Newbury, R.W. and McCullough, G.K.
1982: Shoreline erosion and restabilization in a permafrost-affected impoundment; *in* Proceedings, Fourth International Permafrost Conference, p. 918–921.

Nixon, J.F.
1990: Seasonal and climatic warming effects of pile creep in permafrost; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana, no. 54, p. 335–340.

Nixon, J.F., Sortland, K.A., and James, D.A.
1990: Geotechnical aspects of northern gas pipeline design; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana, no. 54, p. 299–308.

Penner, E. and Crawford, C.B.
1983: Frost action and foundations; Division of Building Research, National Research Council of Canada, DBR Paper no. 1090, 33 p.

Scott, W.J., Sellmann, P.V., and Hunter, J.A.
1990: Geophysics in the study of permafrost; *in* Geotechnical and Environmental Geophysics, v. 1, Review and Tutorial, (ed.) S.H. Ward; Society of Exploration Geophysicists, p. 355–384.

Stanley, D.R.
1996: Water and sewerage problems in discontinuous permafrost regions; *in* Proceedings, Canadian Regional Permafrost conference; National Research Council of Canada, Technical Memorandum no. 86, p. 93–105.

Zarling, J.P., Hansen, P., and Kozisek, L.
1990: Design and performance experience of foundations stabilized with thermosyphons; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana, no. 54, p. 365–370.

Geocryology

Aylsworth, J.M. and Duk-Rodkin, A.
1997: Landslides and permafrost in the Mackenzie Valley; *in* Mackenzie Basin Impact Study (MBIS) Final Report, Environmental Adaptation Research Group; Atmospheric Environment Service, Environment Canada, p. 118–122.

Aylsworth, J.M., Duk-Rodkin, A., Robertson, T., and Traynor, J.A.
in press: Landslides of the Mackenzie valley and adjacent mountainous and coastal regions; *in* The Physical Environment of the Mackenzie River Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547.

Ben-Miloud, K. and Seguin, M.-K.
1990: Stratigraphy, distribution of active layer and discontinuous permafrost in Kangiqsualujjuaq Basin, Northern Quebec; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana, no. 54, p. 131–136.

Bennett, L.P. and French, H.M.
1990: In situ permafrost creep, Melville Island, and implications for global change; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana, no. 54, p. 119–123.

Burgess, M.M. and Smith, S.L.
in press: Shallow ground temperatures in the Mackenzie valley; *in* The Physical Environment of the Mackenzie River Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547.

Dallimore, S.R. and Taylor, A.E.
1993: Permafrost conditions along an onshore-offshore transect of the Canadian Beaufort shelf; *in* Proceedings, Sixth International Permafrost Conference, p. 125–130.

Dredge, L.A., Kerr, D.E., and Wolfe, S.A.
1999: Surficial materials and related ground ice conditions, Slave Province, N.W.T., Canada; Canadian Journal of Earth Sciences, v. 36, p. 1227–1238.

Dyke, L.D.
in press: Stability of permafrost slopes in the Mackenzie valley; *in* The Physical Environment of the Mackenzie River Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada. Bulletin 547.

Dyke, A.S., Morris, T.F., Green, D.E.C., and England, J.
1992: Quaternary geology of Prince of Wales Island, Arctic Canada; Geological Survey of Canada, Memoir 433, 142 p.

French, H.M.
1996: Living on ice: problems of urban development in Canada's north; Geoscience Canada, v. 21, no. 4, p. 163–175.

French, H.M. and Harry, D.G.
1988: Nature and origin of ground ice, Sandhills Moraine, southwest Banks Island, western Canadian Arctic; Journal of Quaternary Science, v. 3, p. 19–30.

French, H.M. and Heginbottom, J.A. (ed.)
1983: Northern Yukon Territory and Mackenzie Delta, Canada. Guidebook to Permafrost and Related Features; Fourth International Conference on Permafrost, Guidebook 3, 186 p.

Hughes, O.L.
1989: Terrain and permafrost: their influence on northern construction; *in* Northern Hydrocarbon Development in the Nineties: a Global Perspective, (ed.) F.T. Frankling; the Geotechnical Science Laboratories, Carleton University, Ottawa, Ontario, p. 109–118.

Judge, A., Pelletier, B.R., and Norquay, I.
1987: Permafrost base and distribution of gas hydrates; *in* Marine Science Atlas of the Beaufort Sea: Geology and Geophysics, (ed.) B.R. Pelletier; Geological Survey of Canada, Miscellaneous Report 40, p. 39.

Lévesque, R., Allard, M., and Seguin, M.K.
1988: Le pergélisol dans les formations quaternaires de la région des rivières Nastapoca et Sheldrake, Québec nordique; Centre d'études nordiques; Université Laval, Québec, Collection Nordicana, n° 51, 23 p.

Lévesque, R., Allard, M., Seguin, M.K., and Pilon, J.-A.
1990: Données préliminaires sur le régime thermique du pergélisol dans quelques localités du Nunavik, Québec; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana, no. 54, p. 207–213.

Lewkowicz, A.G.
1990: Morphology, frequency and magnitude of active-layer detachment slides, Fosheim Peninsula, Ellesmere Island, N.W.T.; *in* Proceedings, Fifth Canadian Permafrost Conference; Université Laval, Quebec, Collection Nordicana, no. 54, p. 111-118.

Lorrain, R.D. and Demeur, P.
1985: Isotopic evidence for relic [*sic*] Pleistocene glacier ice on Victoria Island, Canadian Arctic Archipelago; Arctic and Alpine Research, v. 17, p. 89–98.

Luthin, J.N. and Guymon, G.L.
1974: Soil moisture-vegetation-temperature relationships in central Alaska; Journal of Hydrology, v. 23, p. 233–246.

Mackay, J.R.
1963: The Mackenzie Delta area, Northwest Territories; Department of Mines and Technical Surveys, Geographical Branch, Memoir, 8, 202 p. (reprinted in 1974 as Geological Survey of Canada Miscellaneous Report 23).
1971: The origin of massive icy beds in permafrost, Western Arctic Coast, Canada; Canadian Journal of Earth Sciences, v. 8, p. 397–422.
1972: The world of underground ice; Annals of the Association of American Geographers, v. 62, p. 1–22
1973: Problems in the origin of massive icy beds, western Arctic, Canada; *in* Permafrost: The North American Contribution to the Second International Conference; National Academy of Sciences, Washington, D.C., p. 223–228.

Mackay, J.R. and Dallimore, S.R.
1992: Massive ice of the Tuktoyaktuk area, western Arctic coast, Canada; Canadian Journal of Earth Sciences, v. 29, p. 1235–1249.

Rampton, V.N.
1973: The influence of ground ice and thermokarst upon the geomorphology of the Mackenzie- Beaufort region; *in* Research in Polar and Alpine Geomorphology, (ed.) B.D. Fahey and R.D. Thompson; Proceedings, Third Guelph Symposium on Geomorphology, p. 43–59 .

Seguin, M.K., Allard, M., and Gahe, E.
1989: Surface and downhole geophysics for permafrost mapping in Ungava, Quebec; Physical Geography, v. 10, p. 201–232.

Wolfe, S.A.
1998: Origin of massive ground ice in outwash deposits, northern Slave Province, N.W.T.; Proceedings of Seventh International Conference on Permafrost, Yellowknife, Canada, June 1998; Université Laval, Quebec, Collection Nordicana, no. 57, p. 1133–1140.

Young, S. and Judge, A.S.
1986: Canadian permafrost distribution and thickness data collection: a discussion; *in* Proceedings, National Student Conference on Northern Studies, (ed.) W.P Adams and P.G. Johnson, p. 223–228.

Norman Wells pipeline

Agra Earth and Environmental Ltd. and Nixon Geotech Ltd.
1999: Monograph on Norman Wells Pipeline geotechnical design and performance, Final report to Department of Natural Resources, March 1999; Geological Survey of Canada, Open File 3773, 120 p.

Burgess, M.M. and Tarnocai, C.
1997: Peatlands in the discontinuous permafrost zone along the Norman Wells pipeline, Canada: their characteristics and response to pipeline related environmental changes; *in* Proceedings, International Symposium on Physics, Chemistry and Ecology of Seasonally Frozen Soils, CRREL Special Report 97-10, Alaska, p. 417–424.

Burgess, M.M., Nixon, J.F., and Lawrence, D.E.
1998: Seasonal pipe movement in permafrost terrain, kp2 study site, Norman Wells Pipeline; *in* Proceedings, Seventh International Conference on Permafrost, Yellowknife, June 1998; Université Laval, Quebec. Collection Nordicana, no. 57, p. 95–100.

Hanna, A.J. and McRoberts, E.C.
1988: Permafrost slope design for a buried oil pipeline; *in* Proceedings, Fifth International Permafrost Conference, v. 2, p. 1247–1252.

MacInnes, K.L., Burgess, M.M., Harry, D.G., and Baker, T.H.W.

1989: Permafrost and terrain research and monitoring: Norman Wells Pipeline, v. I Environmental and engineering considerations; Department of Indian and Northern Affairs Canada, Northern Affairs Program, Environmental Studies Report no. 64, 132 p.

1990: Permafrost and terrain research and monitoring: Norman Wells Pipeline, v. II Research and monitoring results: 1983–1988; Department of Indian and Northern Affairs Canada, Northern Affairs Program, 204 p.

Nixon, J.F. and Burgess, M.M.

1999: Norman Wells Pipeline settlement and uplift movement; Canadian Geotechnical Journal, v. 36, p. 119–135.

Climate change

Allard, M., Wang B., and Pilon, J.A.

1995: Recent cooling along the southern shore of Hudson Strait Quebec, Canada, documented from permafrost temperature measurements; Arctic and Alpine Research, v. 27, p. 157–166.

Burgess, M.M., Desrochers, D.T., and Saunders, R.

in press: Potential changes in thaw depth and thaw settlement for three locations in the Mackenzie valley; *in* The Physical Environment of the Mackenzie River Valley, Northwest Territories: a Base Line for the Assessment of Environmental Change, (ed.) L.D. Dyke and G.R. Brooks; Geological Survey of Canada, Bulletin 547.

Burn, C.R.

1998: Field investigations of permafrost and climatic change in northwest North America; *in* Proceedings, Seventh International Conference on Permafrost, Yellowknife, Canada, June 1998; Université Laval, Quebec, Collection Nordicana, no. 57, p. 107–120.

Dyke, L.D., Aylsworth, J.M., Burgess, M.M., Nixon, F.M., and Wright, F.

1997: Permafrost in the Mackenzie Basin, its influences on land-altering processes, and its relationship to climate change; *in* Mackenzie Basin Impact Study (MBIS) Final Report, Environmental Adaptation Research Group, Atmospheric Environment Service, Environment Canada, p. 112–117.

Kettles, I.M., Tarnocai, C., and Bauke, S.D.

1997: Predicted permafrost distribution in Canada under a climate warming scenario; *in* Current Research 1997-E; Geological Survey of Canada, p. 109–115.

Smith, S.L. and Burgess, M.M.

1998: Mapping the response of permafrost in Canada to climate warming; *in* Current Research 1998-E; Geological Survey of Canada, p. 163–171.

Taylor, A.E.

1991: Marine transgression, shoreline emergence: evidence in seabed and terrestrial ground temperatures of changing relative sea levels, Arctic Canada; Journal of Geophysical Research, v. 96, no. B4, p. 6893–6909.

Woo, M., Lewkowicz, A.G., and Rouse, W.R.

1992: Response of the Canadian permafrost environment to climate change; Physical Geography, v. 12, p. 28–43.

Natural gas hydrates

Sharon L. Smith[1]

Smith, S.L., 2001: Natural gas hydrates; in *A Synthesis of Geological Hazards in Canada, (ed.) G.R. Brooks; Geological Survey of Canada, Bulletin 548, p. 265–280.*

Abstract: Temperature and pressure conditions required for the stability of natural gas hydrates are found in sediments of Canada's permafrost and offshore regions. Geophysical evidence suggests that a significant amount of methane is stored as gas hydrate in these regions. Gas hydrates may be an important contributor to future energy supplies and a source of atmospheric greenhouse gas. They may pose a hazard during exploration drilling or hydrocarbon production. Knowledge of their distribution is therefore required to evaluate their importance as an energy resource or drilling hazard. Stability conditions for the existence of gas hydrate in the Canadian permafrost and offshore regions have been analyzed. The distribution of gas hydrate, inferred from geophysical evidence, is described. Examples are presented to illustrate the role climatic and geological history, gas chemistry, and existing temperature and pressure conditions play in determining the stability and distribution of gas hydrate.

Résumé : Les sédiments contenus dans les régions pergélisolées et côtières du Canada présentent des conditions de température et de pression requises pour assurer la stabilité des hydrates de gaz naturel. Les données géophysiques laissent supposer que des quantités significatives de méthane sont stockées dans ces régions sous forme d'hydrates de gaz. Les hydrates de gaz sont susceptibles de fournir d'importants approvisionnements énergétiques dans l'avenir mais constituent également une importante source de gaz à effet de serre. Ils peuvent présenter un risque potentiel lors de l'exécution de forages d'exploration ou de la production d'hydrocarbures. Il est donc indispensable de connaître leur répartition afin d'évaluer leur importance comme ressources énergétiques ou pour estimer les risques qu'ils peuvent faire courir lors de l'exécution de travaux de forage. Les conditions de stabilité indispensables à la présence d'hydrates de gaz dans les régions pergélisolées et extracôtières du Canada ont été analysées. L'auteur décrit la répartition des hydrates de gaz, en se basant sur des données géophysiques. Les exemples présentés permettent d'illustrer le rôle joué par le climat, l'histoire géologique, la chimie des gaz et les conditions de température et de pression sur la détermination de la stabilité et de la répartition des hydrates de gaz.

---

[1] Terrain Sciences Division, Geological Survey of Canada

INTRODUCTION

Natural gas hydrates are ice-like solids that consist of gas molecules, such as methane, enclosed within the cage-like structure of water molecules. The temperature and pressure conditions required for gas hydrates to exist in nature are found in sediments of the permafrost and offshore regions of Canada. Estimates of the amount of methane stored as gas hydrate are highly speculative, but current worldwide estimates are in the order of 2 x 10^{16} m^3. The amount of carbon stored as gas hydrate may be double that present in all known deposits of fossil fuel. Potentially, gas hydrates are an important contributor to future energy supplies and a significant source of atmospheric greenhouse gas.

Because gas hydrate stability is dependent on temperature and pressure, gas hydrate may become unstable if these conditions change. One unit volume of gas hydrate may release 160 volumes of methane (natural gas) during decomposition caused by changes in the temperature or confining pressure of the sediment. Methane, itself a potent greenhouse gas, may be released to the atmosphere because gas hydrates decompose in response to climate warming, and this release of methane may enhance climate warming. On slopes and rises of the outer continental margins, submarine landslides may result from the decomposition of gas hydrate in marine sediments in response to warming of oceaninc bottom waters. During exploration drilling or hydrocarbon production, gas hydrates can become unstable and pose a hazard if proper precautions are not taken. Therefore, a knowledge of the distribution of natural gas hydrates and their potential role in climate change is important; occurrences of natural gas hydrates are directly linked to land and submarine processes and could become unmitigated hazards.

CONDITIONS REQUIRED FOR GAS HYDRATE FORMATION AND STABILITY

Gas hydrates are stable under low temperatures and high pressures. At terrestrial sites, the stability of gas hydrate depends on the surface temperature and the geothermal gradient. As shown in Figure 1, gas hydrate is stable over a greater thickness of sediment in areas where surface temperature is low and the geothermal gradient is small, for example in areas of thick permafrost. Methane hydrate is generally not stable in the upper 200 m of sediment or in areas where permafrost is less than 200 m thick. The presence of a water column or glacier ice influences pressure conditions in the underlying sediment. Conditions for methane hydrate stability at marine sites are illustrated in Figure 2. Gas hydrate stability depends on the depth and temperature of the overlying water and on the geothermal gradient. Methane hydrate is generally not stable beneath water shallower than 200 m deep with a bottom temperature of 0°C or above. In areas covered by glaciers, the thickness of the gas hydrate stability zone increases the thicker the ice cover.

The thickness of the stability zone and therefore the amount of gas that may be stored as gas hydrate will change if temperature or pressure conditions vary. Alterations in surface temperature in response to climate change or fluctuations in sea level, for example, will result in changes in the thickness of the gas hydrate stability zone.

The phase diagrams shown in Figures 1 and 2 represent gas hydrate composed of pure methane and water, which is important to note here because most of the gas in the regions discussed below is methane (*see* 'Hydrate distribution in Canadian permafrost and offshore regions'). Gas chemistry, however, is an important factor that determines the stability of gas hydrates. The addition of other gases, such as hydrogen sulphide or propane, enhances gas hydrate stability and shifts the stability curve to the right; this results in an increase in the thickness of the gas hydrate stability zone. Gas hydrate containing a small amount of hydrogen sulphide, in addition to methane, is stable at shallower depths than if only methane is present.

Temperature and pressure are the main factors determining gas hydrate stability, but the presence of gas hydrate also depends on optimal geological conditions, including the availability of gas and water and the presence of hydrocarbon reservoirs and traps. Typically, decaying biomatter such as peat may serve as a source of gas for gas hydrate. Methane gas from this source is referred to as 'microbial' or 'biogenic' methane. However, methane may also be of thermogenic origin, and form at great depths under high temperatures and pressures. Thermogenic methane can migrate upward along faults or dipping permeable strata until stopped by a relatively impermeable layer such as shale, dolomite, siltstone, permafrost, or existing gas hydrate.

HAZARDS ASSOCIATED WITH NATURAL GAS HYDRATES

Hazards associated with drilling and hydrocarbon production

Natural gas hydrates can pose a serious hazard during exploration drilling and hydrocarbon production. Without proper precautions, uncontrolled gas releases, blowouts, fires, and sediment instability at the well site can occur.

Thermal and mechanical disturbance of gas hydrate during drilling is a common problem. The circulation of warm drilling mud or warm hydrocarbons produced from depth may cause heating of sediments and any gas hydrate contained in them. Gas hydrate can become unstable and dissociate, releasing large amounts of natural gas. As mentioned, volume increases 160-fold when methane stored as gas hydrate becomes a gas at the same temperature and pressure. Free gas trapped around the well casing within the drillhole may produce a substantial increase in reservoir pressures, causing the casing to collapse; also, as gas hydrate decomposes, the surrounding formation may weaken, leading to buckling of the casing. Gas released from gas hydrate formed around the casing may leak to the surface. This is thought to have occurred in the 1970s in the Kuparuk River oil field in Alaska where gas flowed into a rig cellar. Drilling mud may become highly gasified as gas hydrate decomposes and causes severe gas flows (gas kick in well) that may result in fires, as was the case in the Yamburg gas field in western Siberia in the 1960s. An accidental fire at a well on Ellef Rignes Island in the Canadian Arctic in the early 1970s is thought to have been caused by a gas flow produced by decomposing gas hydrate. The use of chilled drilling mud can help reduce these drilling and production hazards. Stresses in the well casing also need to be considered when production is curtailed and gas hydrates begin to reform.

Another drilling hazard is associated with the sealing attributes of gas hydrates, which may trap large volumes of overpressured free gas at shallow depths. Uncontrolled gas releases and well blowouts, such as that reported in the Mackenzie Delta in the 1970s, may result if the gas hydrate trap decomposes.

Gas hydrates may form lenses in sediment, similar to ice lenses, that could present similar geotechnical problems such as thaw settlement or heaving of bottom-founded structures used in offshore oil and gas production, especially where gas hydrates exist at, or close to, the seafloor. The decomposition of gas hydrate due to variations in subsurface sediment temperature in response to climate change or changes in sea level can lead to significant engineering problems such as

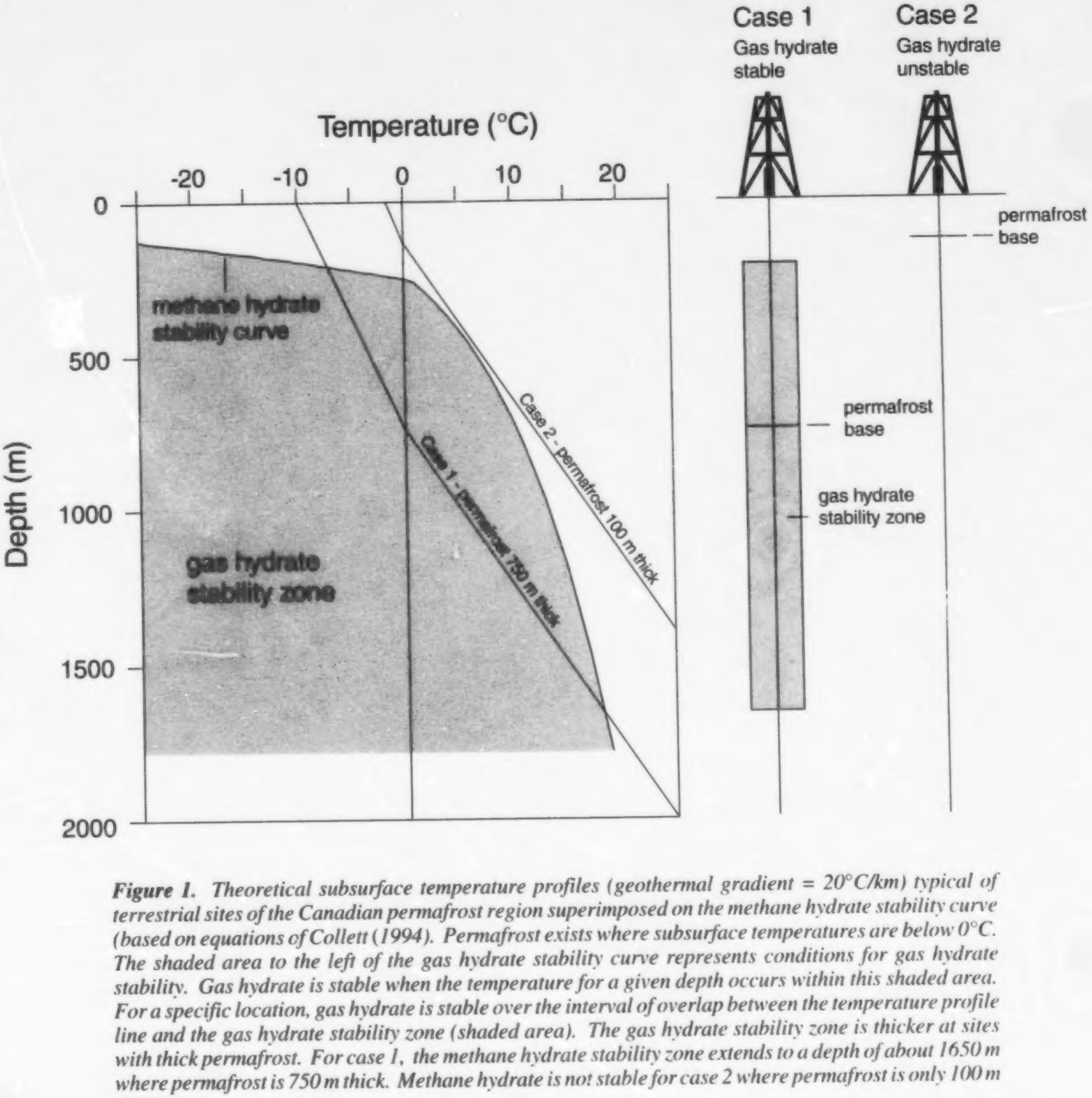


Figure 1. Theoretical subsurface temperature profiles (geothermal gradient = 20°C/km) typical of terrestrial sites of the Canadian permafrost region superimposed on the methane hydrate stability curve (based on equations of Collett (1994). Permafrost exists where subsurface temperatures are below 0°C. The shaded area to the left of the gas hydrate stability curve represents conditions for gas hydrate stability. Gas hydrate is stable when the temperature for a given depth occurs within this shaded area. For a specific location, gas hydrate is stable over the interval of overlap between the temperature profile line and the gas hydrate stability zone (shaded area). The gas hydrate stability zone is thicker at sites with thick permafrost. For case 1, the methane hydrate stability zone extends to a depth of about 1650 m where permafrost is 750 m thick. Methane hydrate is not stable for case 2 where permafrost is only 100 m thick.

underconsolidated sediments, horizons of low strength, or high gas content. These sediments will respond in different ways to natural and induced loads caused by earthquakes, waves, ice, and human activity.

Geological hazard

Gas hydrate can occupy part of the pore space of offshore sediments within the gas hydrate stability zone. Lowering of the sea level or increases in bottom-water temperature can produce conditions in which gas hydrate dissociates at the base of the stability zone (Fig. 3). In addition, continued sedimentation results in deeper burial of gas hydrate, which eventually leads to changes in sediment temperatures at the base of the stability zone. This process creates conditions that are no longer suitable for gas hydrate stability. Although gas hydrate may strengthen sediments within the gas hydrate stability zone, gas hydrate dissociation can cause the release of gas below the gas hydrate stability zone, thus weakening the sediment. Either gravitational loading or seismic disturbances may trigger submarine slope failures through a zone of weakness in the seabed. This action can generate debris flows, slumps, and slides accompanied by a release of methane gas into the water column. Many submarine landslides are thought to have occurred in response to falling sea level associated with the Wisconsinan Glaciation. These slides can be enormous. For example, a 300 km long headwall along the edge of the continental shelf of Norway is associated with a set of three submarine landslides known as the 'Storegga' or 'Great Edge' suite. Seismic evidence reveals that the glide plane of this slide is at the same depth as the base of a gas hydrate layer, indicating that decomposition of gas hydrate resulted in weakening of the sediment. Several other possible

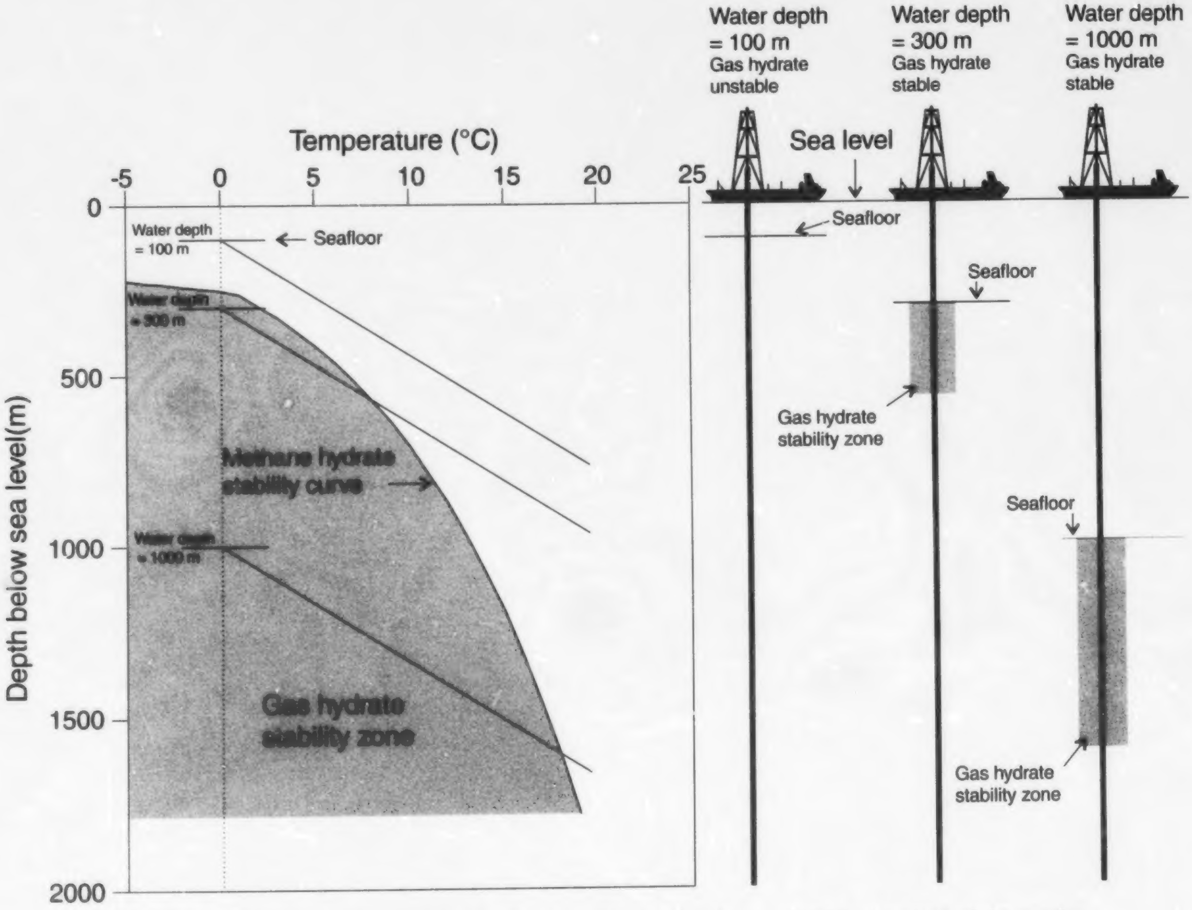


Figure 2. *Theoretical temperature profiles (geothermal gradient = 30°C/km) typical of the marine environment of Canada' East Coast superimposed on the methane hydrate stability curve (based on equations of Collett (1994). The shaded area to the left of the gas hydrate stability curve represents conditions for gas hydrate stability. The interval over which gas hydrate is stable increases with increasing water depth. In the example above, the methane hydrate stability zone is about 250 m thick beneath 300 m of water and 600 m thick beneath 1000 m of water. Methane hydrate is not stable where water is shallow, for example beneath a water depth of 100 m.*

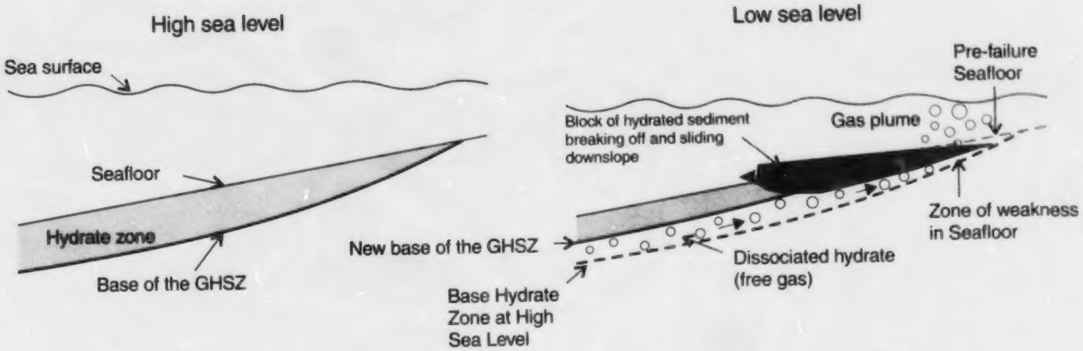

High sea level
Sea surface
Seafloor
Hydrate zone
Base of the GHSZ
Low sea level
Pre-failure Seafloor
Block of hydrated sediment breaking off and sliding downslope
Gas plume
Zone of weakness in Seafloor
New base of the GHSZ
Dissociated hydrate (free gas)
Base Hydrate Zone at High Sea Level

Figure 3. *Gas hydrate dissociation and submarine slope failure following a decrease in pressure caused by a fall in sea level (*modified from *Kvenvolden, 1994). As sea level falls, gas hydrate at the base of the stability zone (GHSZ) dissociates, weakening the strength of the sediment in the sea floor which leads to a landslide and the release of gas.*

examples of submarine slope failures are related to gas hydrate dissociation including several mass movements on the eastern continental margin of the United States, sediment blocks on the seafloor in British Columbia fiords, and massive bedding-plane slides and rotational slumps on the Alaskan part of the Beaufort Sea continental margin. A zone of massive slides and slumps on the Beaufort Slope coincides with a region where sediment is thought to contain gas hydrate. Sea-level lowering related to fluctuations in global climate may have caused these slides, as well as others on various continental margins where gas hydrate is present.

The hazards of submarine landslides include the economic impact of severed fibre-optic submarine cables and pipelines and the generation of massive tsunamis that can have an impact on coastal communities. Transatlantic submarine telegraph cables were cut following a submarine slide (probably related to gas hydrate dissociation) off Newfoundland triggered by the Grand Banks earthquake in 1929. The tsunami associated with this slide caused 27 deaths in Newfoundland. Although catastrophic events are rare, large areas of coastline may be effected by tsunamis that are generated by large submarine landslides.

Gas hydrates and climate change

Methane is a greenhouse gas with a global warming potential that is 20 times that of an equivalent weight of carbon dioxide. The amount of methane trapped in gas hydrate in marine and terrestrial sediments may be 3000 times that currently in the atmosphere, and the partial release of methane from this source in response to climate warming could have a significant effect on the radiative properties of the atmosphere, thus affecting global climate. For example, past changes in global climate during the Pleistocene likely caused the release of methane from gas hydrate, which may have resulted in further modifications of the global climate. The transgression of the polar ocean over the exposed, colder continental shelf as sea level rose during deglaciation resulted in an increase in sediment temperature and the destabilization of gas hydrate. Although the melting of ice sheets in response to climate warming caused the sea level to rise and stabilize submarine gas hydrate in the seabed, continental gas hydrates became unstable as air temperatures increased.

During a climatic cooling cycle ice sheets grow, causing the sea level to fall. As the polar continental shelves are exposed and subjected to low subaerial temperatures and become colder, gas hydrate can form in the newly emerged land. Conditions become favourable for gas hydrate stability in continental regions as ice sheets advance and pressure in the underlying sediments increases. Submarine gas hydrates within sediments of continental slopes and rises, however, become unstable due to a decrease in pressure as the sea level drops. During global warming and cooling cycles associated with interglacial and glacial climates, gas hydrate deposits should respond to changes in surface pressure and temperature.

Although the effects of methane from gas hydrates on global climate are uncertain, theories have been proposed to explain the role of gas hydrate in past global climate change. It has been suggested, for example, that methane from the decomposition of polar continental and continental shelf gas hydrates contributed to the rapid rise in atmospheric methane, carbon dioxide (oxidation of methane in the atmosphere results in the formation of carbon dioxide) and global temperature at the end of the Wisconsinan Glaciation about 13 500 years ago. Increased temperature and decreased pressure in the subsurface that accompanied the melting of ice sheets would lead to the destabilization of gas hydrate and the emission of methane. The increase in atmospheric methane would lead to further warming, which in turn would cause further gas hydrate decomposition and methane emission, possibly helping to end the ice age.

Another scenario suggests that deep-sea gas hydrate deposits in the outer continental margins released methane as sea level dropped during the buildup of the Wisconsinan ice sheets. Methane was possibly released to the atmosphere in sudden bursts as submarine slumping occured and gas was released abruptly. The addition of this methane to the atmosphere initiated climatic warming, which triggered the beginning of deglaciation. Methane derived from gas hydrate deposits in the outer continental margin may, therefore, have been a factor that limited the duration and extent of glaciation during a glacial cycle. Although there is evidence that large submarine slope failures occurred during the Wisconsinan, evidence for huge, sudden methane bursts is indirect.

These ideas are speculative and, in fact, the behaviour of gas hydrate in the existing climate regime is unknown. Climatic warming is expected to be greater in the polar regions under current global warming scenarios; therefore, gas hydrates in the Arctic are most vulnerable to climate change. Modelling results indicate that, although submarine hydrates will remain stable for the next 1000 years as water temperatures increase, gas hydrate in the permafrost regions could begin decomposing in the next 100 years under catastrophic warming scenarios of an increase in temperature of 8°C/100 a. If a more conservative increase of 0.6°C/100 a is considered, the release of methane would be substantially slower. Of note is the report of in situ gas hydrate occurring above the methane hydrate stability zone. This gas hydrate could be affected by warming over a shorter time period.

Gas hydrate deposits of the polar continental shelves such as those beneath the Beaufort Sea are probably the most vulnerable to climate change. These areas were formerly exposed to very cold surface temperatures and have been, and are being, transgressed by a much warmer polar ocean. This has resulted in a temperature increase of over 10°C in the past 10 000 years. A further rise in sea level in response to climate warming and the accompanying inundation of coastal areas in this region would result in an increase in the area containing unstable gas hydrate.

HYDRATE DISTRIBUTION IN CANADIAN PERMAFROST AND OFFSHORE REGIONS

The distribution of natural gas hydrate is summarized for the following five regions (Fig. 4; Northern Geothermal Consultants, unpub. rept., 1991, 1993; Thurber Consultants Ltd.,

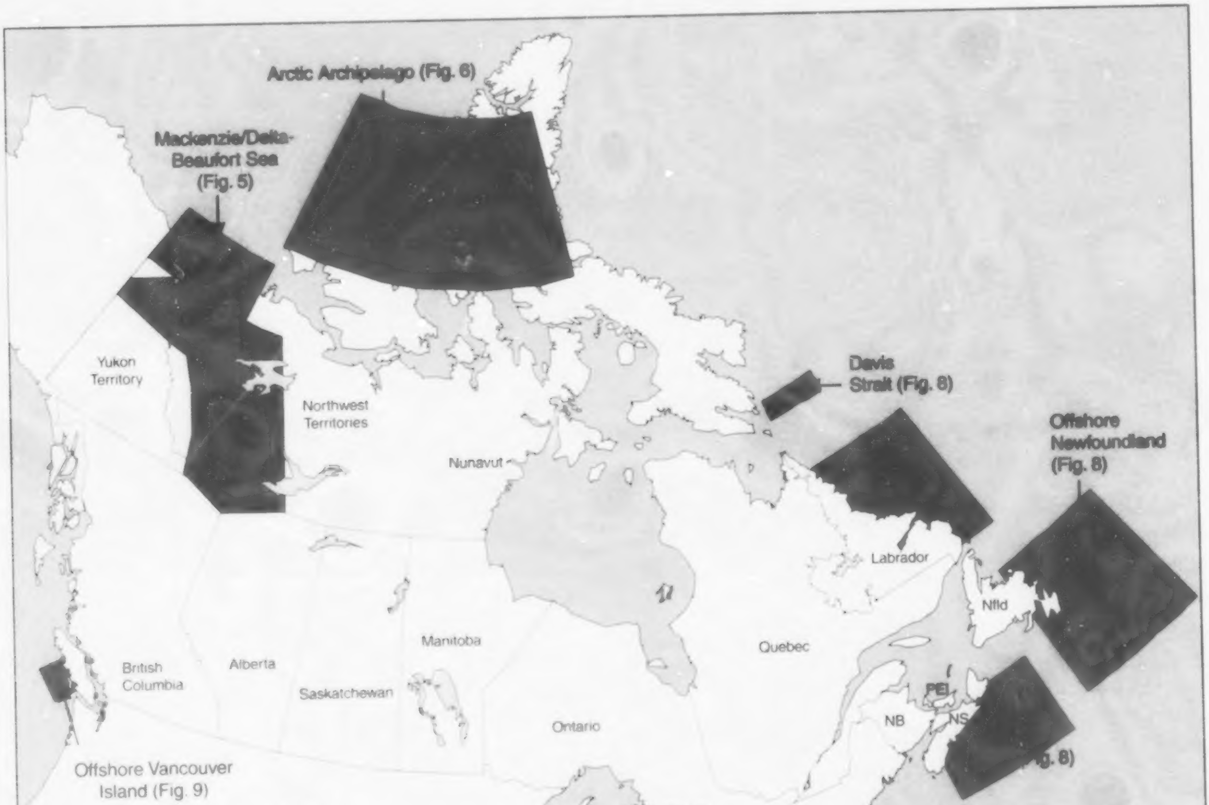

Arctic Archipelago (Fig. 6)
Mackenzie/Delta-Beaufort Sea (Fig. 5)
Yukon Territory
Northwest Territories
Nunavut
Davis Strait (Fig. 8)
Offshore Newfoundland (Fig. 8)
Labrador
Nfld
Quebec
Manitoba
British Columbia
Alberta
Saskatchewan
Ontario
PEI
NB
NS
g. 8)
Offshore Vancouver Island (Fig. 9)
0 km 1000

Figure 4. Regions in Canada where the distribution and stability of natural gas hydrates are reviewed in the text. Locations referred to in Figures 5, 6, 7, 8, and 9 are shown.

unpub. rept., 1985): Mackenzie Delta and Beaufort Sea (Fig. 5), the islands and channels of the Canadian Arctic Archipelago (Fig. 6), the western Northwest Territories and the Yukon Territory (Fig. 7), the offshore region of Eastern Canada (Fig. 8), and the offshore region of Western Canada (Fig. 9). Figures 5 to 9 show areas where conditions exist for gas hydrate stability and where evidence has been found for the existence of gas hydrates.

Determination of the gas hydrate stability zone

Deep-temperature data and information on the depth to the base of ice-bearing permafrost from about 500 hydrocarbon exploration wells and precise temperature surveys in the upper few hundred metres were used to outline that part of the permafrost region of Canada where methane hydrate may be stable. Ground-temperature profiles are superimposed on the methane hydrate stability curve as illustrated in Figure 1 to determine if conditions exist for methane hydrate stability. Information on seabottom temperature (water temperature and depth data were provided by the Marine Environmental Data Services, Department of Fisheries and Oceans), water depth, and geothermal gradient was used to identify (using the method illustrated in Figure 2) where conditions for gas hydrate stability are found in the offshore regions of Eastern Canada.

In determining the extent of the gas hydrate stability zone, only the stability of methane hydrate was considered because most of the gas in the regions in question is of this type. The

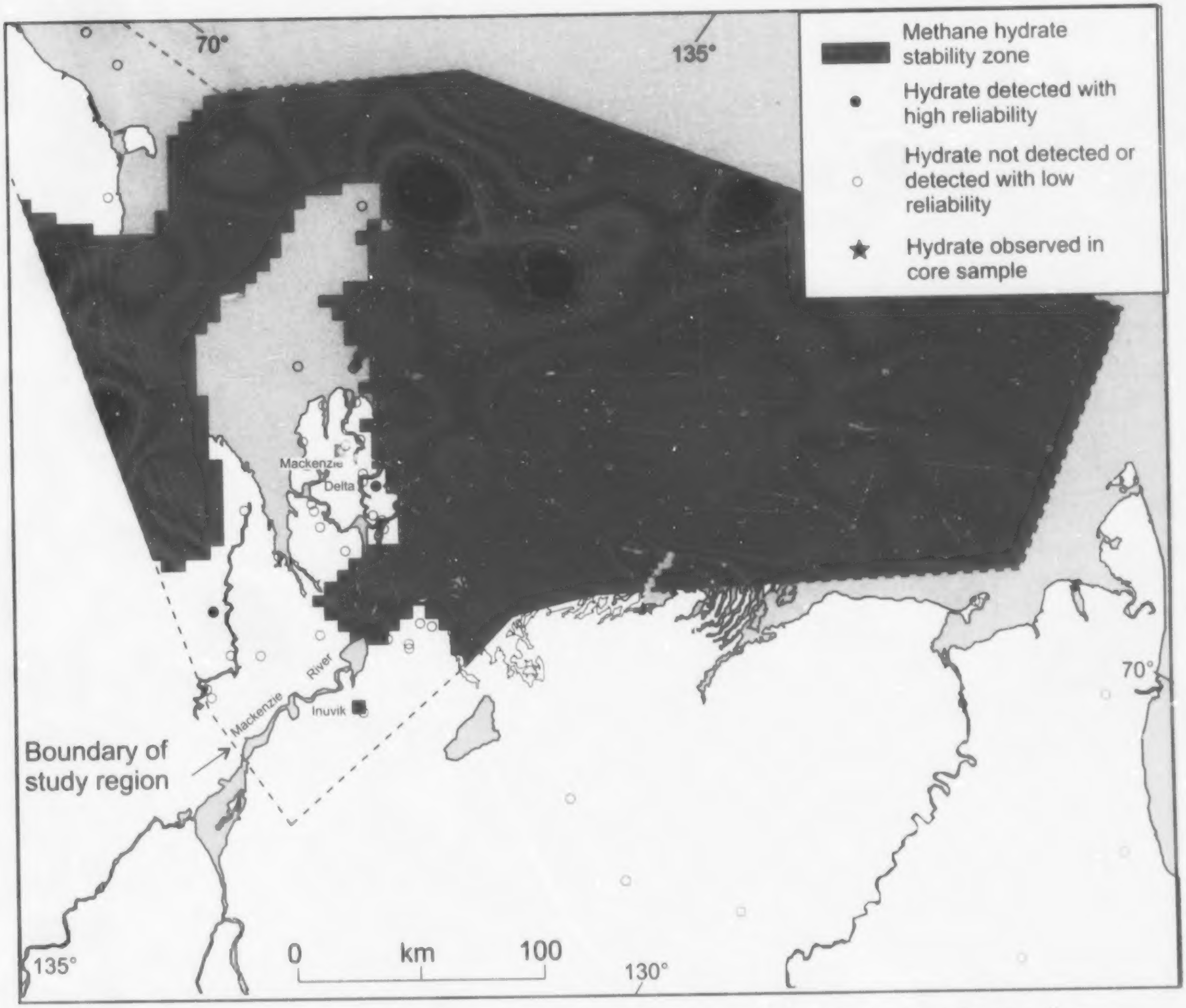

Methane hydrate stability zone
● Hydrate detected with high reliability
○ Hydrate not detected or detected with low reliability
★ Hydrate observed in core sample
70°
135°
Mackenzie Delta
Mackenzie River
Inuvik
Boundary of study region
70°
0 km 100
135°
130°

Figure 5. Areal extent of the methane hydrate stability zone in the Mackenzie Delta–Beaufort Sea region. Also shown are the locations of wells where gas hydrate has been detected with high reliability and sites where gas hydrate was either not detected or detected with low reliability. The boundary of the study region delineates the area for which temperature data were available to determine hydrate stability.

presence of other gases, such as heavier hydrocarbons or hydrogen sulphide, may enhance gas hydrate stability; in some areas, the gas hydrate stability zone may be more extensive than that shown on the maps.

Geophysical evidence for gas hydrate

Petrophysical data from over 500 exploratory oil and gas wells were examined to see whether or not gas hydrate was present in the permafrost and eastern offshore regions (geophysical evidence for gas hydrates is summarized by Smith and Judge (1993)). Petrophysical characteristics (dependent to some extent on drilling conditions) of gas-hydrate-bearing zones may include high electrical resistivity, very low sonic velocity with cycle skipping, low density, high neutron porosity, and enlarged borehole diameter on the caliper log. The presence of gas hydrate may be further confirmed with mud gas logs, where available.

In offshore regions, information from seismic surveys can be used to infer the presence of gas hydrate. The appearance of an anomalous acoustic reflector cutting across bedding-plane reflectors on seismic reflection profiles is considered an indicator of gas hydrates within sediments of the continental margin. This anomalous reflection is called a 'bottom-simulating reflector' or 'BSR'. Another indicator of possible gas hydrate occurrence observed on seismic reflection profiles is the 'bright spot' that develops where an accumulation of free gas is present below the gas hydrate zone.

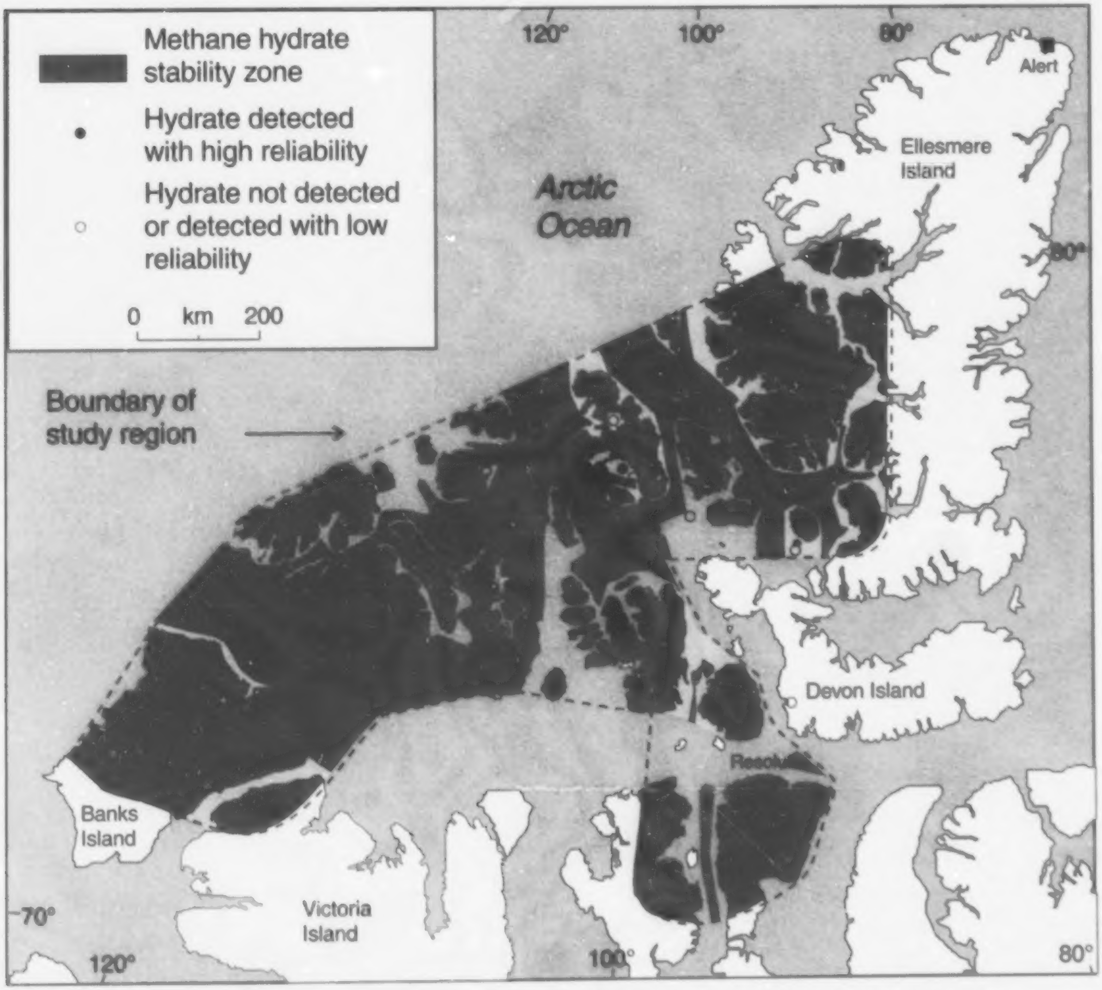
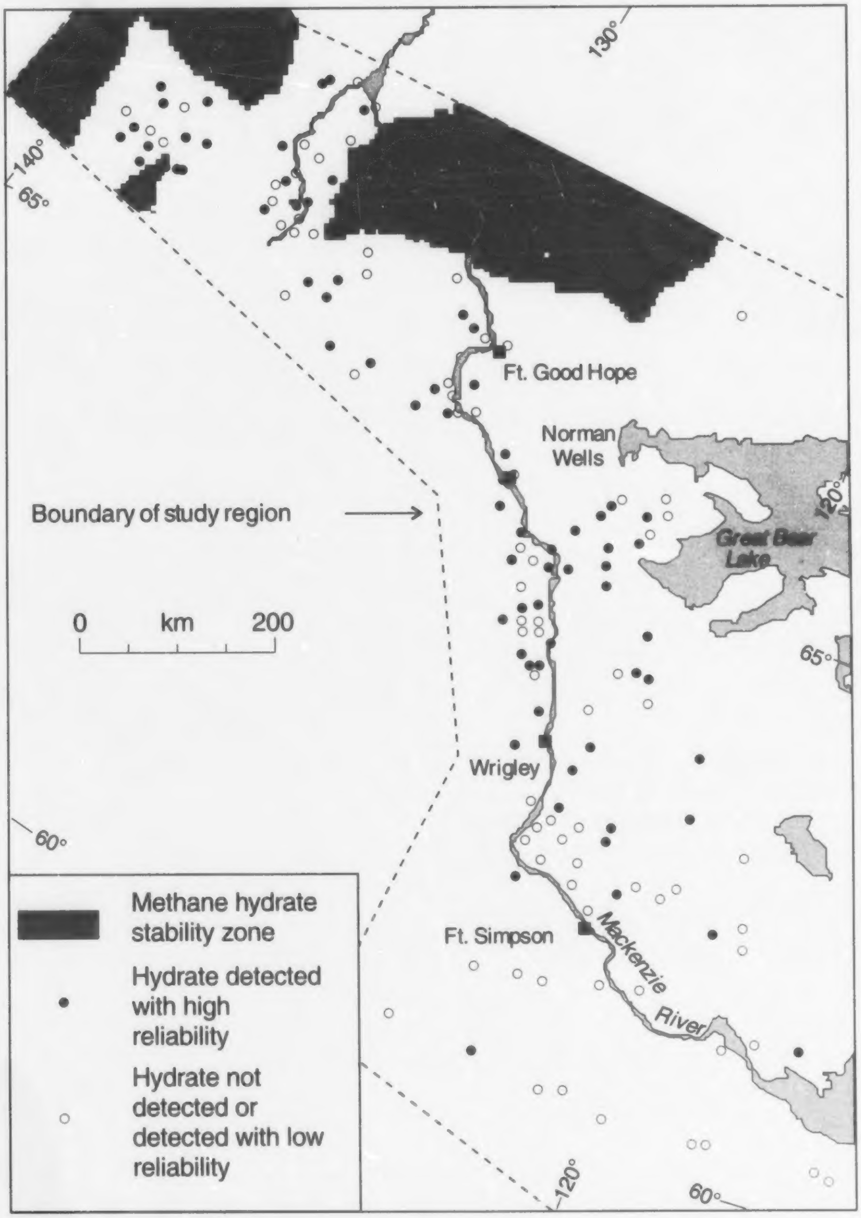

Methane hydrate stability zone
Hydrate detected with high reliability
Hydrate not detected or detected with low reliability
0 km 200
Boundary of study region
Arctic Ocean
120°
100°
80°
Alert
Ellesmere Island
80°
Devon Island
Resolute
Banks Island
Victoria Island
70°
120°
100°
80°

Figure 6. Areal extent of the methane hydrate stability zone in the Canadian Arctic Archipelago. Also shown are the locations of wells where gas hydrate has been detected with high reliability and sites where gas hydrate was either not detected or detected with low reliability. The boundary of the study region delineates the area for which temperature data were available in order to determine hydrate stability.

Mackenzie Delta–Beaufort Sea region

An extensive methane-hydrate-prone zone extending to depths as great as 1400 m exists in the Mackenzie Delta–Beaufort Sea region (Fig. 5). Gas hydrate is generally unstable in sediments beneath the modern Mackenzie Delta where permafrost is thin. The methane hydrate stability zone is thickest in the sediments of Richards Island and Tuktoyaktuk Peninsula where thick permafrost exists. Conditions for methane hydrate stability are also found beneath the Beaufort Shelf east of latitude 136°W. This area remained



130°
140°
65°
Ft. Good Hope
Norman Wells
Boundary of study region
Great Bear Lake
120°
0 km 200
65°
Wrigley
60°
Ft. Simpson
Mackenzie River
120°
60°
Methane hydrate stability zone
Hydrate detected with high reliability
Hydrate not detected or detected with low reliability

Figure 7. Areal extent of the methane hydrate stability zone in the Yukon Territory and western Northwest Territories. Also shown are the locations of wells where gas hydrate has been detected with high reliability and sites where gas hydrate was either not detected or detected with low reliability. The boundary of the study region delineates the area for which temperature data were available to determine hydrate stability.

ice-free when the sea level fell 140 m during the middle Wisconsinan. Sediments were exposed to much lower temperatures than sediments that were covered by ice or water, such as those of the western Beaufort Shelf. A thick permafrost and gas-hydrate-prone layer formed beneath the eastern Beaufort Shelf. Over the past 27 000 years, however, a rise in sea level has led to warming and erosion of the gas hydrate stability zone in this area (Fig. 10).

Gas hydrate has been detected with high reliability in 14% of terrestrial wells and 45% of offshore wells (Fig. 5); it has also been inferred from seismic profiles beneath the Beaufort Shelf. It is less common at terrestrial sites where hydrocarbon reservoirs are generally located in formations below the gas hydrate stability zone. At marine sites, gas has migrated along faults into the younger formations found within the gas hydrate stability zone. Gas hydrate has been observed within ice-bonded permafrost in core samples from a site on Richards Island (Fig. 5). This is thought to be the first confirmed occurrence of gas hydrate within permafrost. Ice-bearing permafrost and gas hydrate exhibit a similar petrophysical response, making it difficult to detect gas hydrate within permafrost; this suggests that significant accumulations of intra-permafrost gas hydrate may have been overlooked because they cannot be readily distinguished from ice-bearing sediments. Pore-space gas hydrate was also observed at depths of 119 m, which suggests that gas hydrate may exist at depths shallower than the stability zone predicted for methane hydrate. Thus, gas hydrate may be present within permafrost in the area shown in Figure 10, although it has not been detected. Geophysical evidence also indicates that gas

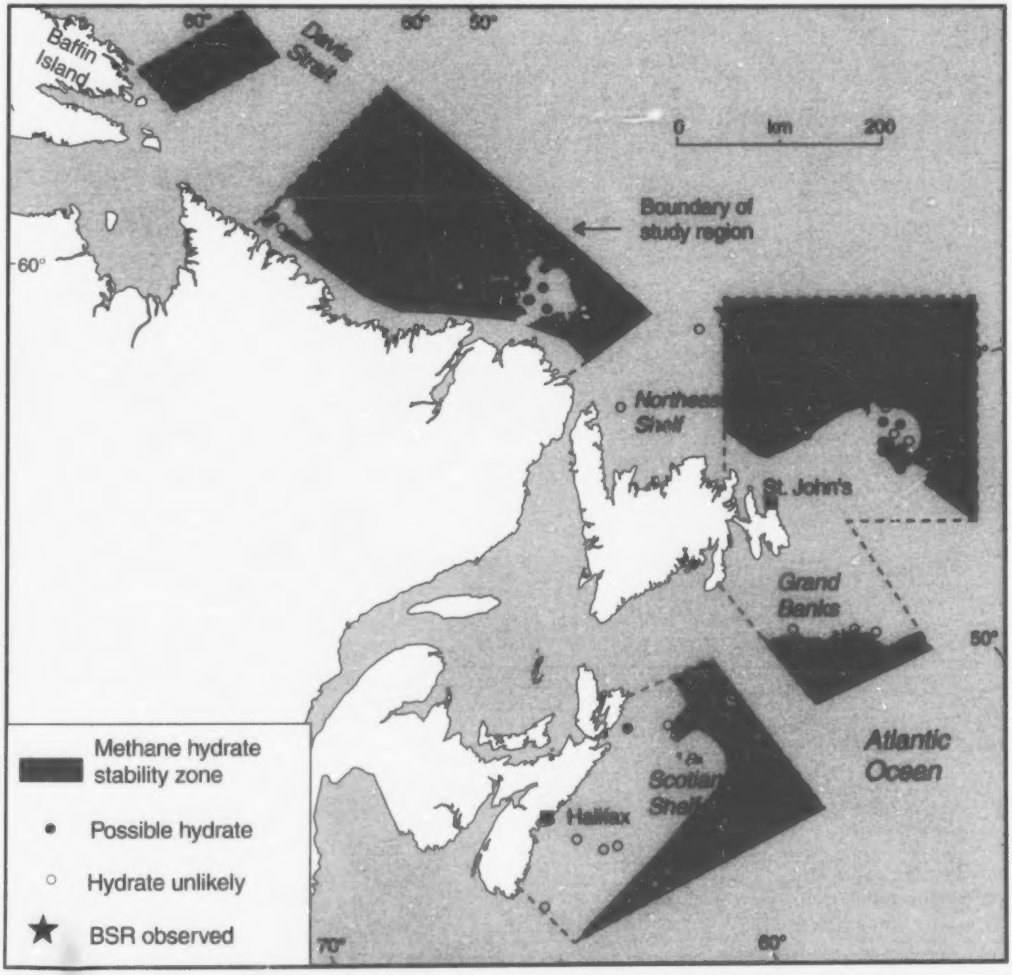


Figure 8. Areal extent of the methane hydrate stability zone in the offshore region of Eastern Canada. Also shown are the locations of wells where gas hydrate has been interpreted from well logs to possibly exist and sites where gas hydrate is unlikely to occur. Sites where bottom-simulating reflectors have been observed are also shown. The boundary of the study region delineates the area for which temperature data were available to determine hydrate stability.

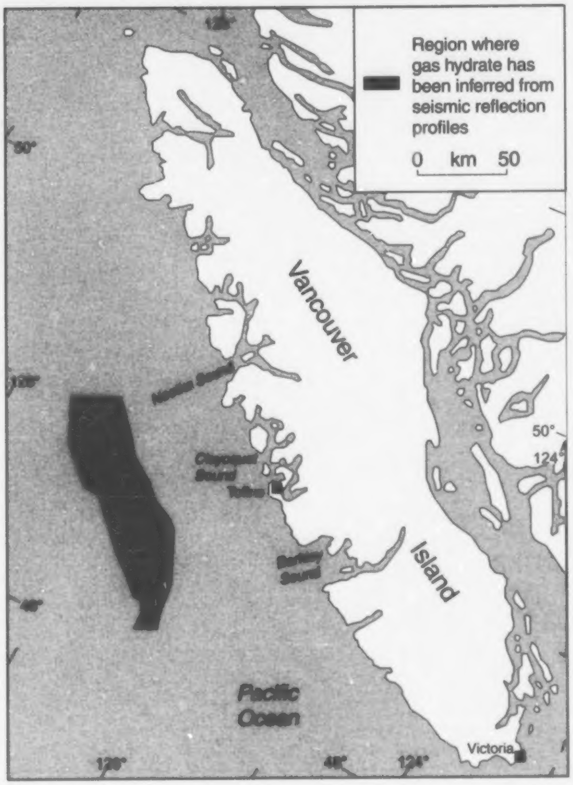


*Figure 9. Area off the western coast of Vancouver Island where seismic evidence (bottom-simulating reflectors) indicates the presence of gas hydrate (*modified from *Hyndman, 1992). The hydrate zone extends along the mid-continental slope off the coast of Washington and Oregon.*

hydrate may exist outside the methane hydrate stability zone (Fig. 10) and this may be related to the presence of heavier hydrocarbons.

Gas hydrates beneath the Beaufort Shelf will become unstable and decompose as subsurface warming continues in response to marine transgression. The decomposition of gas hydrate may have contributed to the formation of gassy sediments found in some areas. Its effect on the geotechnical properties of sediments could have important implications for the development of offshore oil and gas. For example, the loss of sediment strength due to gas hydrate decomposition would need to be considered in designing bottom-founded structures. Sea level continues to rise (0.1–0.12 cm/a) and the further inundation of coastal areas will increase the area containing an unstable gas hydrate zone. Over a 100 year period, any gas hydrate in the upper 2 to 3 m of the methane hydrate stability zone would become unstable.

Canadian Arctic Archipelago

In large areas of the channels and islands of the Canadian Arctic Archipelago, conditions are suitable for the occurrence of methane hydrate (Fig. 6). An extensive methane-hydrate-prone layer exists beneath these islands to depths of 2000 m and in the intervening channels to depths of 1200 m. Gas hydrate has been detected with high reliability in 24% of terrestrial wells and 23% of marine wells. Geophysical evidence suggests that gas hydrate is present mostly in the western Sverdrup Basin, where major gas fields are also located. In the eastern part of the basin, gas hydrate is less common, possibly because potential hydrocarbon reservoirs were breached during the Eurekan Orogeny.

The extent of the gas-hydrate-prone zone is inconsistent in this region and has been changing and continues to change over time. The Sverdrup Basin has been undergoing postglacial uplift over the last 1000 years at an average rate of between 0.3 and 0.5 m every 100 years. As the environment of the emerging land changes from marine to terrestrial in coastal regions, the ground surface cools by 15°C. This cooling results in permafrost aggradation and creates conditions for gas hydrate stability, as illustrated in Figure 11 for the coastal region of the Sabine Peninsula of Melville Island. The area currently below an elevation of 60 m has emerged from the ocean over the past 15 000 years. Areas at higher elevations have thicker permafrost and gas-hydrate-prone zones than those at lower elevations. Close to the shoreline where the ground surface has been exposed to the lower temperatures for only a relatively short time, existing conditions are probably unsuitable for gas hydrate stability. The coastal region acts as a methane sink (storage) and will continue to do so as long as emergence and permafrost growth continue. Gas hydrate is possibly present within the stability zone at the sites shown in Figure 11. Natural gas has also been discovered in this area (near well 73), which suggests that hydrocarbons may be available for the formation of gas hydrate.

Although gas hydrate may be forming in coastal regions, methane hydrate is unstable in offshore areas where water depths are less than 200 m. As uplift occurs, relative sea level falls and gas hydrate that was stable under greater water depths becomes unstable. About 37% of the area of the Sverdrup Basin seafloor is shallower than 200 m and is subject to degassing. These areas may be a source of atmospheric methane.

The possible existence of a thick ice sheet covering most of the Sverdrup Basin during the Wisconsinan Glaciation may have allowed the development of a thick gas-hydrate-prone zone in areas where temperature and pressure conditions are currently unsuitable for hydrate formation. Gas hydrate that formed in sediments beneath the ice sheet would decompose as pressure decreased during deglaciation, and this process may still be going on in marine sediments beneath shallow water. For example, gas storages and gas vents inferred to exist from acoustic data in the sediments of Expedition Fiord, Axel Heiberg Island, may have been produced as a result of degrading gas hydrate. During glaciation,

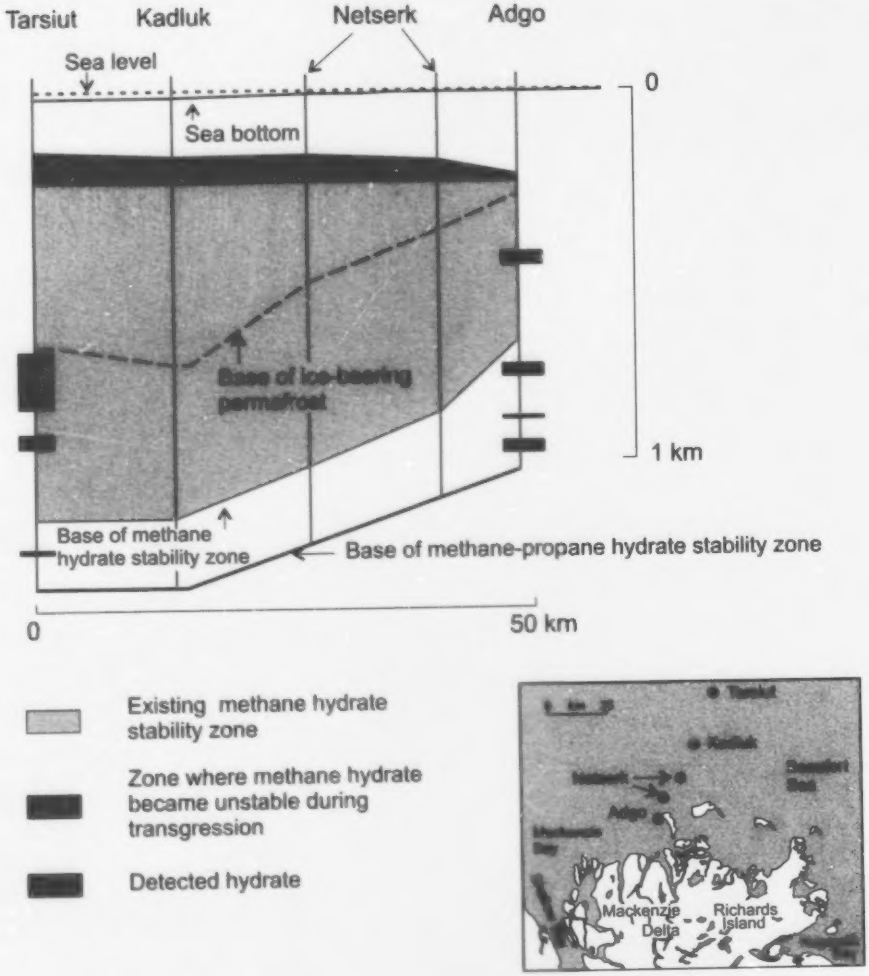

Tarsiut
Kadluk
Netserk
Adgo
Sea level
0
Sea bottom
Base of ice-bearing permafrost
1 km
Base of methane hydrate stability zone
Base of methane-propane hydrate stability zone
0
50 km
Existing methane hydrate stability zone
Zone where methane hydrate became unstable during transgression
Detected hydrate
Tarsiut
Kadluk
Netserk
Adgo
Mackenzie Delta
Richards Island

Figure 10. Transect of the methane hydrate and methane-propane hydrate stability zones below the Beaufort Sea shelf north of the Mackenzie Delta at approximately longitude 136°W (inset shows well locations). The error associated with the calculation of the base of the gas hydrate stability zone is ±200 m. This area was exposed to cold, subaerial temperatures during the Laurentide Glaciation, which allowed a thick permafrost and gas-hydrate-prone layer to form. Subsequent sea-level rise led to inundation of the sites and warming of sediments. The period of warming has been longest at the Tarsiut site and shortest at the Adgo site where the water depth is shallower. The degradation of the upper part of the methane hydrate stability zone is also shown, and any gas hydrate that may have existed within this zone has become unstable during transgression. The erosion of the upper part of the gas hydrate stability zone is greater farther offshore (Tarsiut site), where the shelf sediments have been warming for a longer period (Judge et al., 1994).

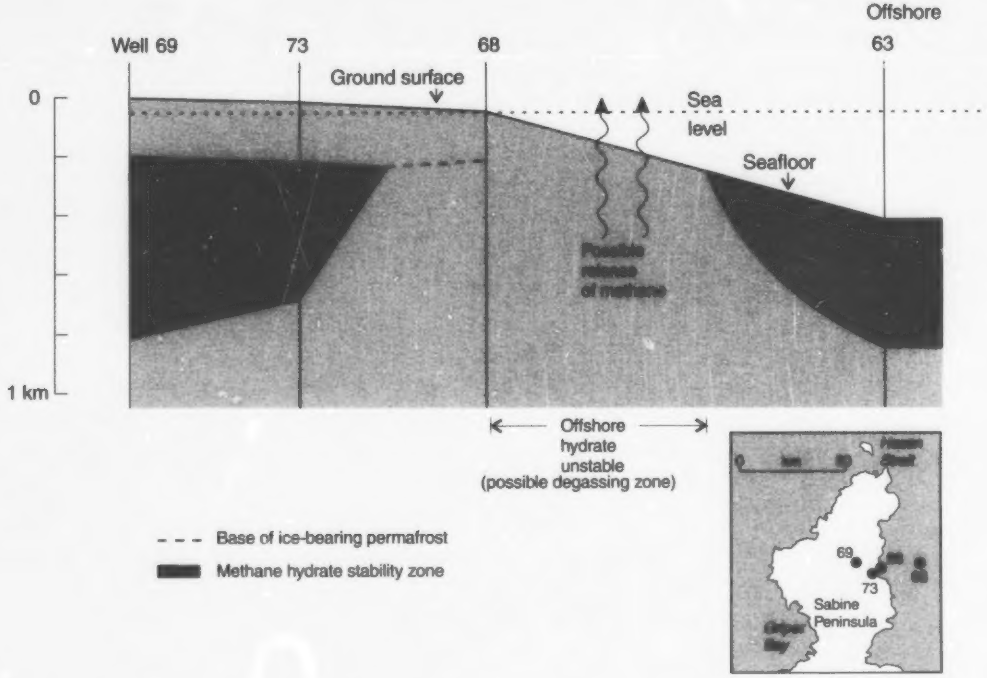

Offshore
Well 69
73
68
63
Ground surface
Sea level
Seafloor
0
1 km
Possible release of methane
Offshore hydrate unstable (possible degassing zone)
- - - Base of ice-bearing permafrost
Methane hydrate stability zone
0
69
73
Sabine Peninsula

Figure 11. Transect of the coastal area on the east side of the Sabine Peninsula, Melville Island (well locations shown in inset). There is no horizontal scale on the diagram. The permafrost and gas hydrate stability zone thickens with increasing elevation and distance landward (from well 68 to 69) because the period of exposure to lower subaerial temperature (and thus cooling) is longer. Methane hydrate is also stable beneath the deeper water of the island channels (well 63).

ice several hundreds of metres thick, may have been present in the fiords, allowing the formation of gas hydrate in these areas. Currently, water depths are only 40 to 50 m and conditions are unsuitable for methane hydrate stability.

Western Northwest Territories—Yukon Territory

Methane hydrate is unstable over about 89% of the western Northwest Territories—Yukon Territory region (Fig. 7) where permafrost is generally thin. Conditions for methane hydrate stability are found in the northern part of the region where geothermal gradients are lower and permafrost is thicker. At its maximum thickness, the methane-hydrate-prone zone extends to a depth of about 830 m. The presence of gas hydrate has been interpreted from well logs as probable at 38% of the well sites (Fig. 7). It has been inferred with good reliability at depths as great as 2000 m and as shallow as 15 m.

Gas chemistry is an important factor to consider in accounting for the many inferred occurrences of gas hydrate outside the methane hydrate stability zone (Fig. 7). Gas hydrate in this region may contain propane and other heavier hydrocarbons such as butane, which form gas hydrate at higher temperatures and lower pressures than methane hydrate. Hydrogen sulphide has also been reported in some well logs and this may also enhance the stability of methane hydrate. Gas hydrate has been interpreted to occur at depths shallower than 200 m and, as mentioned earlier, gas hydrate composed of these other gases is stable at shallower depths than methane hydrate. Therefore, gas composition may play an important role in determining the thickness and extent of the stability zone in this region.

Another factor to consider is the role of glaciation in the distribution of gas hydrate in this region. During the Wisconsinan Glaciation, an extensive methane-hydrate-prone layer would have existed under a thick ice cover and large volumes of methane gas may have accumulated in the form of gas hydrate at shallow depths. Some of this gas hydrate may not have decomposed following the pressure decrease associated with glacial retreat and may still be present today. Shallow gas hydrates may be 'self-preserved' gas hydrates that are unstable, relict gas hydrates. Self-preserved gas hydrates have been observed in the laboratory and may occur in nature as well.

The possible existence of shallow gas hydrates in this region and the hazard they may pose must be considered during seismic and geotechnical investigations and mining exploration, as well as during hydrocarbon exploration and development. Shallow gas hydrate may become unstable over the next 50 to 100 years under existing climate-warming scenarios and release methane gas to the atmosphere.

Eastern Canada

In the offshore region of Eastern Canada, methane hydrate is generally stable only in areas where water is over 200 m deep. Extensive methane-hydrate-prone layers are present in the Davis Strait and Labrador Shelf regions (Fig. 8). Methane hydrate is stable below the deeper waters of the northeastern Newfoundland Shelf and unstable in the shallow waters of the Grand Banks. Water depths are generally too shallow and water temperatures too warm for methane hydrate stability on the Scotian Shelf. Conditions for methane hydrate stability, however, are found in the deep-water areas of the Scotian Slope and Grand Banks Slope (areas offshore from the Scotian Shelf and Grand Banks).

Gas hydrate has been interpreted to be possibly present in 75% of the wells examined (Fig. 8). Geophysical evidence for gas hydrate occurrence has been found in Davis Strait and the Labrador Shelf. A possible bottom-simulating reflector provides additional evidence for the presence of gas hydrate beneath the Labrador Shelf. Gas hydrate has been interpreted to exist from well logs beneath the northeastern Newfoundland Shelf and from the identification of bottom-simulating reflectors from seismic data farther offshore in the Orphan Basin. Limited geophysical evidence exists for gas hydrate at sites on the Grand Banks and Scotian Shelf. Geophysical data from well logs indicate that gas hydrate may exist where water depths are too shallow for methane hydrate stability (Fig. 8). Relict gas hydrate that formed beneath areas formerly covered by Pleistocene ice or exposed to colder air temperatures during periods of lower sea level may be present in this region.

Western Canada

Conditions for methane hydrate stability also exist beneath the continental slope west of Vancouver Island where water depths are greater than 400 m. Bottom-simulating reflectors have been observed on seismic profiles and are widespread in a 30 km wide zone beneath much of the continental slope (Fig. 9). The gas hydrate layer is interpreted to be 20 m thick and is found at the base of the methane hydrate stability zone, approximately 300 m below the seafloor. Seismic evidence indicates that gas hydrate is not present where water depths are greater than about 1000 m.

SUMMARY

Conditions for methane hydrate stability exist in the permafrost and offshore regions of Canada. Data from geophysical well logs and seismic profiles indicate that gas hydrate is present in these areas. Some of the areas in which gas hydrates may pose a hazard are shown in Figures 4 to 9. It is important to note, however, that only the stability of methane hydrate has been considered in the production of the maps. The presence of other gases may lead to the formation of gas hydrate over a more extensive area. A second point to consider is that the existence of gas hydrate is generally inferred from geophysical evidence, rather than based on direct observation. It is also difficult to distinguish between gas hydrate and ice in the permafrost region. Commonly, mud gas logs provide the only evidence of gas hydrate in these sediments. Thus, significant amounts of gas hydrate could be present within permafrost and have been overlooked because of the difficulty in detecting their presence. This has important implications for hazard assessment and climate-change studies in the permafrost regions.

The potential extent of the marine gas hydrate zone off the East and West coasts has not been fully explored in this contribution. The amount of gas hydrate is much greater off the West Coast, but the concern for gas hydrate hazard is most important off the East Coast because of current petroleum exploration and development in areas such as the deep-water zones of the Scotian Slope and Grand Banks Slope where conditions for gas hydrate stability are found.

Examples presented here show that the extent of the gas hydrate stability zone and the existence of gas hydrate depend on climatic and geological history, gas chemistry, and existing temperature and pressure conditions. Thus, any assessment of the hazards related to gas hydrate must take these factors into account. For example, gas hydrate may be present at shallower depths than those predicted for methane hydrate stability, and this unanticipated occurrence of gas hydrate may pose a significant hazard during drilling. The factors influencing the stability and distribution of gas hydrate also need to be considered when estimating the amount of methane that may be released in response to climate warming and the effect this release may have on climate in the future.

ACKNOWLEDGMENTS

This analysis depended on high-quality measurements made during exploratory drilling in northern Canada, and the dedication of numerous oil industry, logging company, and government regulatory personnel is acknowledged. Daniel T. Desrochers of Environmental Analysis Services assisted in the production of the maps. Comments on the manuscript by Greg Brooks, Bernard Pelletier, Scott Dallimore, David Piper, and Roy Hyndman of the Geological Survey of Canada are greatly appreciated.

REFERENCES

Collett, T.S.
1994: Permafrost-associated gas hydrate accumulations; *in* International Conference on Natural Gas Hydrates, (ed.) E.D. Sloan, J. Happel, and M.A. Hnatow; Annals of the New York Academy of Sciences, v. 715, p. 247–269.

Hyndman, R.D.
1992: "Ice" beneath the deep sea: studies of methane hydrate layers beneath the continental slope off Vancouver Island; Geoscience Canada, v. 19, p. 21–26.

Judge, A., Smith, S.L., and Majorowicz, J.
1994: The current distribution and thermal stability of natural gas hydrates in the Canadian polar regions; *in* Proceedings, Fourth (1994) International Offshore and Polar Engineering Conference, Osaka, Japan, v. 1, p. 307–314.

Kvenvolden, K.A.
1994: Hydrate occurrence; *in* International Conference on Natural Gas Hydrates, (ed.) E.D. Sloan, J. Happel, and M.A. Hnatow; Annals of the New York Academy of Sciences, v. 715, p. 232–246.

Smith, S.L. and Judge, A.S.
1993: Gas hydrate database for Canadian Arctic and selected East Coast wells; Geological Survey of Canada, Open File 2746, 120 p.

SUGGESTED READING

General reviews and conference proceedings

Collett, T.S., Bird, J.J., Kvenvolden, K.A., and Magoon, L.B.
1988: Geologic interrelations relative to gas hydrates within the North Slope of Alaska; United States Geological Survey, Open File 88-389, 150 p.

Cox, J.L. (ed.)
1983: Natural Gas Hydrates: Properties, Occurrences, and Recovery; Butterworth Publishers, Boston, Massachussetts, 125 p.

Davidson, D.W.
1973: Clathrate hydrates; *in* Water: A Comprehensive Treatise, (ed.) F. Franks; Plenum Press, New York, v. 2, p. 115–234.

Kvenvolden, K.A.
1982: Occurrence and origin of marine gas hydrates; *in* Proceedings, Fourth Canadian Permafrost Conference, p. 305–311.

Kvenvolden, K.A., Ginsburg, G.D., and Soloviev, V.A.
1993: Worldwide distribution of subaquatic gas hydrates; Geomarine Letters, v. 13, p. 32–40.

Sloan, E.D.
1990: Clathrate Hydrates of Natural Gases; Marcel Dekker, New York, 641 p.

Gas hydrates in Canada

Bily, C. and Dick, J.W.L.
1974: Naturally occurring gas hydrates in the Mackenzie Delta, N.W.T.; Bulletin of Canadian Petroleum Geology, v. 22, 340–353.

Carson, B., Westbrook, G.K., Musgrave, R., and Suess, E. (ed.)
1995: Proceedings of ODP, Scientific Report, 146 (Part 1); College Station, Texas, Ocean Drilling Program, 477 p.

Collett, T.S. and Dallimore, S.R.
1998: Quantitative assessment of gas hydrates in the Mallik 2L-38 well, Mackenzie Delta, N.W.T.; *in* Proceedings, Seventh International Permafrost Conference; Université Laval, Collection Nordicana, no. 57, p. 189–194.

Dallimore, S.R. and Collett, T.S.
1995: Intra permafrost gas hydrates from a deep core hole in the Mackenzie Delta, Northwest Territories, Canada; Geology, v. 23, p. 527–530.
1998: Gas hydrates associated with deep permafrost in the Mackenzie Delta, N.W.T., Canada: regional overview; *in* Proceedings, Seventh International Permafrost Conference; Université Laval, Collection Nordicana, no. 57, p. 201–206.

Davidson, D.W., El-Defrawy, M.K., Feuglem, M.O., and Judge, A.S.
1978: Natural gas hydrates in northern Canada; *in* Proceedings, Third International Conference on Permafrost, v. 1, p. 937–943.

Judge, A.
1982: Natural gas hydrates in Canada; *in* Proceedings, Fourth Canadian Permafrost Conference, p. 320–328.

Judge, A.S. and Majorowicz, J.A.
1992: Geothermal conditions for gas hydrate stability in the Beaufort-Mackenzie area: The global change aspect; Global and Planetary Change, v. 6, p. 251–263.

Judge, A.S., Jones, I.G., and Lewis, C.F.M.
1990: Gas hydrates; *in* Chapter 14 of Geology of the Continental Margin of Eastern Canada, (ed.) M.J. Keen and G.L. Williams; Geological Survey of Canada, Geology of Canada, no. 2, p. 771–774 (*also* Geological Society of American, The Geology of North America, v. I-1, p. 771–774).

Judge, A.S., Pelletier, B.R., and Norquay, I. in Pelletier, B.R. (ed.)
1987: Permafrost base and distribution of gas hydrates; *in* Marine Science Atlas of the Beaufort Sea, Geology and Geophysics, (ed.) B.R. Pelletier; Geological Survey of Canada, Miscellaneous Report 40, 39 p.

Smith, S.L.
1998: An analysis of the thermal field to determine constraints on gas hydrate stability in Yukon Territory and western Northwest Territories; *in* Current Research 1998-B; Geological Survey of Canada, p. 235–241.

Taylor, A.E., Wetmiller, R.J., and Judge, A.S.
1979: Two risks to drilling and production off the east coast of Canada — earthquakes and gas hydrates; *in* Proceedings, Symposium on Research in the Labrador Coastal and Offshore Region, (ed.) W. Denner; Memorial University of Newfoundland, p. 91–105.

Weaver, J.S. and Stewart, J.M.
1982: In situ hydrates under the Beaufort shelf; *in* Proceedings, Fourth Canadian Permafrost Conference, p. 312–319.

Drilling and geological hazard

Collett, T.S.
1990: Potential geologic hazards of arctic gas hydrates; American Association of Petroleum Geologists, v. 74, p. 631–632.

Nisbet, E.G. and Piper, D.J.W.
1998: Giant submarine landslides; Nature, v. 392, p. 329–330.

Rothwell, R.G., Thomson, J., and Kahler, G.
1998: Low-sea-level emplacement of a very large Late Pleistocene 'megaturbidite' in the western Mediterranean Sea; Nature, v. 392, p. 377–380.

Yakushev, V.S. and Collett, T.S.
1992: Gas hydrates in arctic regions: risk to drilling and production; *in* Proceedings, Second (1992) International Offshore and Polar Engineering Conference, San Francisco, v. 1, p. 669–673.

Climate change and gas hydrate

Kvenvolden, K.A.
1988: Methane hydrates and global climate; Global Biogeochemical Cycles, v. 2, p. 221–229.
1991: A review of arctic gas hydrates as a source of methane in global change; *in* International Conference on the Role of the Polar Regions in Climate Change, (ed.) G. Weller, C.L. Wilson, and B.A.B. Severin; v. II, p. 696–701.

MacDonald, G.J.
1990: Role of methane clathrates in past and future climates; Climatic Change, v. 16, p. 247–281.

Nisbet, E.G.
1989: Some northern sources of atmospheric methane; production, history, and future implications; Canadian Journal of Earth Science, v. 26, p. 1603–1611.

Paull, C.K., Ussler, W., and Dillon, W.P.
1991: Is the extent of glaciation limited by marine gas-hydrates; Geophysical Research Letters, v. 18, p. 432–434.

Sea-level change, paleoclimate, and glaciology

Allen, D., Michel, F., and Judge, A.
1988: Paleoclimate and permafrost in the Mackenzie Delta; *in* Proceedings, Fifth International Permafrost Conference, v. 1, p. 33–38

Andrews, J.T. and Peltier, W.R.
1989: Quaternary geodynamics in Canada; Chapter 8 *in* Quaternary Geology of Canada and Greenland, (ed.) R.J. Fulton; Geological Survey of Canada, Geology of Canada, no. 1, p. 543–572 (*also* Geological Society of America, The Geology of North America, v. K-1, p. 543–572).

Gilbert, R., Aitken, A.E., and Lemmen, D.S.
1993: The glacimarine sedimentary environment of Expedition Fiord, Canadian High Arctic; Marine-Geology, v. 110, p. 257–273.

Hill, P.R., Hequette, A., and Ruz, M-H
1993: Holocene sea-level history of the Canadian Beaufort Shelf; Canadian Journal of Earth Science, v. 30, p. 103–108.

Hill, P.R., Mudie, P.J., Moran, K., and Blasco, S.M.
1985: A sea-level curve for the Canadian Beaufort Shelf; Canadian Journal of Earth Science, v. 22, p. 1381–1398.

McLaren, P. and Barnett, D.M.
1978: Holocene emergence of the south and east coasts of Melville Island, Queen Elizabeth Islands, Northwest Territories, Canada; Arctic, v. 31, p. 415–427.

Piper, D.J.W., Mudie, P.J., Fader, G.B., Josenhans, H.W., MacLean, B., and Vilks, G.
1990: Quaternary geology; Chapter 10 *in* Geology of the Continental Margin of Eastern Canada, (ed.) M.J. Keen and G.L. Williams; Geological Survey of Canada, Geology of Canada, no. 2, p. 475–607 (*also* Geological Society of America, The Geology of North America, v. I-1, p. 475–607).

King, L.H. and Fader, G.B.J.
1986: Wisconsinan glaciation of the Atlantic Continental shelf of southeast Canada; Geological Survey of Canada, Bulletin 363, 72 p.

Taylor, A.E.
1991: Marine transgression, shoreline emergence: evidence in seabed and terrestrial ground temperatures of changing relative sea levels, Arctic Canada; Journal of Geophysical Research, v. 96, p. 6893–6909.

Heat flow

Issler, D.R.
1984: Calculation of organic maturation levels for offshore eastern Canada — implications for general application of Lopatin's method; Canadian Journal of Earth Sciences, v. 21, p. 477–488.

Lewis, J.F. and Hyndman, R.D.
1976: Oceanic heat flow measurements over the continental margins of eastern Canada; Canadian Journal of Earth Sciences, v. 13, p. 1031–1038.

Pye, G.D. and Hyndman, R.D.
1972: Heat flow measurements in Baffin Bay and the Labrador Sea; Journal of Geophysical Research, v. 77, p. 938–944.

Reiter, M. and Jessop, A.M.
1985: Estimates of terrestrial heat flow in offshore eastern Canada; Canadian Journal of Earth Sciences, v. 22, p. 1503–1517.

Additional data sources used to produce Figures 4 to 9

D&S Petrophysical Consultants
1983: A study of well logs in the Mackenzie Delta/Beaufort Sea to outline permafrost and gas hydrate occurrence; Energy, Mines, and Resources Canada, Earth Physics Branch, Open File 83-10, 242 p.

GEOTECH Ltd.
1983: Subsurface temperature data from Arctic wells; Energy, Mines, and Resources Canada, Earth Physics Branch, Open File 83-11.
1984: Subsurface temperature data from wells north of sixty, Yukon-Northwest Territories; Energy, Mines, and Resources Canada, Earth Physics Branch, Open File 84-28, 557 p.

Hardy Associates (1978) Ltd.
1984: A study of well logs in the Western Northwest Territories and Yukon to outline permafrost thickness and/or gas hydrate occurrence; Energy, Mines, and Resources Canada, Earth Physics Branch, Open File 84-27, 290 p.
1984: Study of well logs in the Arctic Islands to outline permafrost thickness and/or gas hydrate occurrence (two volumes); Energy, Mines, and Resources Canada, Earth Physics Branch, Open File 84-8.

Taylor, A.E., Burgess, M., Judge, A.S., and Allen, V.S.
1982: Canadian geothermal data collection — northern wells 1981; Energy, Mines, and Resources Canada, Earth Physics Branch, Geothermal Series no. 13, 153 p.

AUTHOR INDEX

Adams, J. 7
Basham, P. 7
Boteler, D.H. 183
Brooks, G.R. 1, 101
Burgess, M.M. 241
Clague, J.J. 27, 101
Edwards, B.R. 145
Evans, S.G. 43, 101
Forbes, D.L. 225
Grieve, R.A.F. 207
Heginottom, J.A. 241
Hickson, C.J. 145
Jamieson, B. 81
Klassen, R.A. 227
Knight, R.D. 227
Shaw, J. 225
Smith, S.L. 241, 265
Solomon, S. 225
Taylor, R.B. 225
Wolfe, S.A. 231